PARTS OF THE ORIGINAL DOCUMENT

ARE OF POOR REPRODUCTIVE QUALITY

CERTAINES PARTIES DU DOCUMENT ORIGINAL

SONT DE PIÈTRE QUALITÉ DE REPRODUCTION

Syllogeus 55

Climatic Change in Canada 5

Critical Periods in the Quaternary Climatic History of Northern North America

C.R. Harington, Editor

National Museums of Canada
Musées nationaux du Canada
National Museum of Natural Sciences
Musée national des sciences naturelles

88-03282/

SYLLOGEUS is a publication of the National Museum of Natural Sciences, National Museums of Canada, designed to permit the rapid dissemination of information pertaining to those disciplines and educational functions for which the National Museum of Natural Sciences is responsible. In the interests of making information available quickly, normal publishing procedures have been abbreviated.

Articles are published in English, in French, or in both languages, and the issues appear at irregular intervals. A complete list of the titles issued since the beginning of the series (1972) and individual copies of this number are available by mail from the National Museum of Natural Sciences, Ottawa, Canada K1A 0M8.

La collection SYLLOGEUS, publiée par le Musée national des sciences naturelles, Musées nationaux du Canada, a pour but de diffuser rapidement le résultat des travaux dans les domaines scientifique et éducatif qui sont sous la direction du Musée national des sciences naturelles. Pour assurer la prompte distribution de cette publication, on a abregé les étapes de la rédaction.

Les articles sont publiés en français, en anglais ou dans les deux langues, et ils paraissent irrégulièrement. On peut obtenir par commande postale la liste des titres de tous les articles publiés depuis le début de la collection (1972) et des copies individuelles de ce numéro, au Musée national des sciences naturelles, Ottawa, Canada K1A 0M8.

Syllogeus Series No. 55
(c) National Museums of Canada 1985
Printed in Canada
ISSN 0704-576X

Serie Syllogeus No. 55
(c) Musées nationaux du Canada 1985
Imprimé au Canada

CLIMATIC CHANGE IN CANADA 5

Ye Hudson's Bay Co
1816
AUGUST
S M T W T F S

"THE YEAR WITHOUT A SUMMER" AT A HUDSON'S BAY COMPANY TRADING POST.

INSIDE COVER CARTOON BY CHARLES DOUGLAS

CLIMATIC CHANGE IN CANADA 5

CRITICAL PERIODS IN THE QUATERNARY CLIMATIC HISTORY OF NORTHERN NORTH AMERICA

National Museum of Natural Sciences
Project on Climatic Change in Canada
During the Past 20,000 Years

Edited by C.R. HARINGTON

Syllogeus No. 55

National Museums of Canada
National Museum of Natural Sciences

Les Musées nationaux du Canada
Musée national des sciences naturelles

ACKNOWLEDGEMENTS

The editor is grateful to: C.G. Gruchy and Ridgeley Williams (then Acting Director and Acting Assistant Director, Research and Operations, National Museum of Natural Sciences, respectively) for their support; colleagues in the Paleobiology Division for their continuing interest and encouragement of the project; Helen Davies (Special Projects Officer, NMNS), who provided great help organizing and supervising catering; Marie-Thérèse D'Amour (Public Relations Officer, NMNS) for spreading news of the meeting to the public and arranging for the information folders provided to participants; Eleanor Fenton (Publications Coordinator, NMNS) for help in publication of the Program and Abstracts booklet; Charles Diotte (Foreman of Building Maintenance, NMNS) and his crew for setting up the Salon of the Victoria Memorial Museum Building, where the symposium was held; Marcel Demers and Richard Martin for audiovisual arrangements; and the volunteers, Judy Murillo and Cécile Rymek, for their help at the registration desk. The editor particularly wishes to thank Gail Rice (Assistant to the Division Chief, Paleobiology Division) for invaluable assistance in organizing the meeting and for help in preparation of this volume.

In 1984, a grant was received via Atmospheric Environment Service, which had been allocated extra resources to develop specific areas of the Canadian Climate Program, including work undertaken by the NMNS Climatic Change Project. The editor sincerely thanks Howard Ferguson, John Sandilands and Mal Berry of the Canadian Climate Centre for their assistance.

CONTENTS

Introduction

C.R. Harington 6

Avant-Propos

C.R. Harington 15

Special Lecture

The Value of Paleoclimatic Data to a Climatologist

M.K. Thomas 27

Bibliographies

Holocene Paleoclimates: an Annotated Bibliography

Martha Andrews 31

Annotated Bibliography of Quaternary Climatic Change in Canada, and a Brief Analysis of Its Contents

C.R. Harington 53

Instrumental Records

Thirties Drought on the Prairies - How Unique Was It?

M.O. Berry and G.D.V. Williams 63

The Effects of Major Volcanic Eruptions on Canadian Climate

Walter R. Skinner 75

Historic Records

Sea Ice and Climatic Change in the Canadian Arctic Since 1800

Moira Dunbar 107

Evidence from Hudson Bay Region of Severe Cold in the Summer of 1816

A.J.W. Catchpole 121

The Little Ice Age on Eastern Hudson/James Bay: the Summer Weather and Climate at Great Whale, Fort George and Eastmain, 1814-1821, as Derived from Hudson's Bay Company Records

Cynthia Wilson 147

Daily Weather Maps for Canada, Summers 1816 to 1818 - a Pilot Study

Cynthia Wilson 191

A Dramatic Change in the General Circulation on the West Coast of Hudson Bay in 1760 A.D.: Synoptic Evidence Based on Historic Records

Timothy Ball 219

A Reconstruction of New England Climate Using Historical Materials, 1620-1980

William R. Baron and Geoffrey A. Gordon 229

Prehistory

Prehistoric Cultural Distributions as an Indicator of Environmental Change
(Abstract)

J.V. Wright 247

Tree Rings

Investigating the Possibility of a Relationship Between Volcanic Eruptions and Tree Growth in Canada (1800-1899)

M.L. Parker 249

Summer Degree-Days Since 1574 in Northwestern Canada and Alaska (Abstract)

G.C. Jacoby, Jr., L.D. Ulan and E.R. Cook 265

Fossil Pollen

The Postglacial Development of Vegetation in Newfoundland and Eastern Labrador-Ungava: Synthesis and Climatic Implications

Joyce Brown Macpherson 267

Late-Glacial Climatic Change in the Maritime Provinces

R.J. Mott 281

Mean July Temperature at 6000 yr B.P. in Eastern North America: Regression Equations for Estimates from Fossil-Pollen Data

Patrick J. Bartlein and Thompson Webb, III 301

Postglacial Pollen and Paleoclimate in Southwestern Alberta and Southeastern British Columbia

L.V. Hills et al.

Part 1: Palynologic and Paleoclimatic Interpretation of Holocene Sediments, Waterton Lakes National Park, Alberta

O.A. Christensen and L.V. Hills 345

Part 2: A Palynological Record, Upper Elk Valley, British Columbia

Angus Fergusson and L.V. Hills 355

Part 3: Holocene Palynology of Crowsnest Lake, Alberta, with Comments on Holocene Paleoenvironments of the Southern Alberta Rockies and Surrounding Areas

J.C. Driver, L.V. Hills and B.O.K. Reeves 370

Paleobotanical Evidence for Climatic Change in Southern British Columbia During Late-Glacial and Holocene Time

Rolf W. Mathewes 397

Glacial Geology and Geochronology

Reconstruction of Environmental Conditions in the Eastern Canadian Arctic During the Last 11,000 Years

John T. Andrews 423

Old Crow Tephra: Its Significance for Understanding the Early and Middle Wisconsinan Climate History of Eastern Beringia

John V. Matthews, Jr. and Charles E. Schweger 453

Paleoclimatology and Glaciology

1550-1620: a Period of Summer Accumulation in the Queen Elizabeth Islands

Bea Taylor Alt 461

How Did We Get into the Last Glaciation, How Did We Get Out of It, and What Happened in Between?
(Abstract)

C.U. Hammer 481

INTRODUCTION

C.R. Harington[1]

Since its inception in 1977, the National Museum of Natural Sciences Climatic Change Project held three meetings attended by members and associates. This meeting (May 19-20, 1983, held in the Victoria Memorial Museum Building, Ottawa) was our fourth, and the Project's first international meeting. Its aim was, by focusing on certain "critical periods" in the Quaternary climatic history of northern North America (e.g. the postglacial melting about 10,000 years ago, the Hypsithermal, the Little Climatic Optimum about 1000 A.D., the Little Ice Age about 1500-1850 A.D., times of major volcanic eruptions and years of extreme cold, drought, etc.), to clarify their nature, causes and impacts as viewed from various fields.

Of particular note, was the presentation at the beginning of the meeting of a plaque to Morley K. Thomas for his outstanding contribution to the study of climatic change in Canada. Ridgeley Williams made the presentation on behalf of the NMNS Climatic Change Project.

Morley first encountered meteorology as a student at the University of Western Ontario. He spent four years as a forecaster with the Royal Canadian Air Force. At the close of the Second World War in 1945, he studied climatology at Toronto, receiving a Masters degree in 1948-49. He then moved through the ranks in the federal government, first as Superintendent of Operations in the Climate Division; in 1971 becoming Director of the Meteorological Applications Branch; in 1976 becoming Director-General of Central Services; and in 1979-80 taking up a special assignment as Director-General of the Canadian Climate Centre in Downsview prior to his retirement. He is now working on a book covering the history of meteorology in Canada.

Morley has published many scientific and popular papers, as well as books on climatology. He has participated in numerous international and regional climatology conferences, and from 1978-82 headed the Commission for Climatology and Applications of Meteorology - participating in planning of both the World Climate Program and the Canadian Climate Program. His main professional interests are in the fields of applications and climatic change. In this light, it is easy to understand the bold, practical message he imparts in the Special Lecture, which followed presentation of the plaque: ... "paleoclimatic data will become increasingly important to climatologists and hence to society, both as classical and practical knowledge. The data are a most essential ingredient to our knowledge of past and future climates. I know that your work is valuable and important and so do you, but we have to convince a wide circle of people. *You* can be the most valuable salesman or promoter your work can ever have".

---

[1] Paleobiology Division, National Museum of Natural Sciences, National Museums of Canada, Ottawa, Ontario, K1A 0M8

The contents of this symposium volume are organized, somewhat arbitrarily, under headings denoting the main discipline or disciplines involved: Bibliographies; Instrumental Records; Historic Records; Prehistory; Tree Rings; Fossil Pollen; Glacial Geology and Geochronology; and Paleoclimatology and Glaciology. Most papers are based on Late-Glacial and Holocene fossil pollen analysis (seven), and on studies of historical records (six) - particularly Hudson's Bay Company archives. As far as time is concerned, nearly half of the papers deal with the Holocene (particularly focusing on the Hypsithermal), whereas about a quarter cover the 1800s - several concentrating on "the year without a summer" (1816). Geographically, the papers deal with all regions of Canada, particularly Hudson Bay and the Arctic. New England, Alaska and Greenland are also mentioned.

Bibliographies are useful tools for workers trying to gain background information on critical periods in the Quaternary climatic history of northern North America. This volume contains two papers on the subject. The first, by Martha Andrews, describes the process of selecting, organizing and computerizing data for use in the "Annotated Bibliography of Holocene Paleoclimates". She notes that her results are preliminary. The second paper, by C.R. Harington, provides information on the "Annotated Bibliography of Quaternary Climatic Change in Canada". Although differences in categorization can be seen (e.g. 16 subject categories in the former compared to nearly 40 in the latter, and nearly twice the number of geographic descriptors for Canada in the former compared to the latter), some of the conclusions are strikingly similar (e.g. the relatively high number of papers dealing with glacial geology and fossil pollen, and the surprising quantity of papers referring to the Canadian Arctic Islands). Of course, the differences in scope of the two bibliographies must be recognized. The first focuses on approximately the last 10,000 years - the scale being worldwide, whereas the second spans about the last two million years, narrowing its geographic coverage to Canada.

Perhaps the most critical climatic event affecting Canadian society during the past century was the 1930s drought on the Prairies. Berry and Williams indicate that such events are not uncommon. Indeed, conditions as dry as the worst year of the 1930s may be expected with an average frequency of once in 20 years. This clearly demonstrates the need for careful contingency planning in order to ameliorate such critical periods in future.

Have major volcanic eruptions, since that of Krakatau in 1883, had an impact on Canadian climate? Walter Skinner sheds some sorely needed light on this problem. Using the relatively sensitive superposed epoch analysis method developed by Mass and Schneider in 1977, he indicates that there were marked annual decreases in Canadian temperatures (and to a lesser extent, precipitation) after major eruptions. He finds that the temperature decrease in the Arctic was slightly greater than for the country as a whole, amounting to a drop of about $1^{\circ}C$ during both the eruption year and the following year. Further, a drop of about 13 mm in Prairie precipitation was detected during the growing season (May - September) after mid-latitude eruptions. It is worth comparing this work with Marion Parker's paper, which sets out to determine if expected decreases in temperature following the eruptions of Tambora in 1815 and Krakatau are reflected in the Canadian tree-ring record. Parker detects no continent-wide effects using the technique of dendroclimatology, but admits that further studies are required. Skinner's results support the theoretical results of Pollack *et al.* (1976) and the empirical findings of Mass and Schneider (1977) and Taylor (1978), but not the condensation nuclei hypothesis of Wexler (1951) suggesting increased precipitation following major volcanic eruptions.

Moira Dunbar, using historical records of explorers and whalers, makes the case that ice conditions were more severe during the peak period of exploration in the Canadian Arctic Islands (about 1818-1860) than in this century. Such evidence should be carefully considered by historians in their evaluation of the success, or lack of it, of expeditions searching for Sir John Franklin.

Hudson's Bay Company archives are a treasure house of information on climate in central and northern Canada during the last few hundred years. Papers by Alan Catchpole, Cynthia Wilson and Timothy Ball show how useful this information can be in outlining the nature of weather and climate in the above regions, particularly during the 1800s.

Alan Catchpole discusses "the year without a summer", 1816, in the course of his work showing how substantial the Company's records for the Bay region are for this period: nineteen daily post journals, nine daily temperature records, four ships' log books and a journal kept on a coastal sloop! He makes interesting use of Cri Lake, Québec tree-ring data (Parker *et al.* 1981) in order to calculate mean summer temperatures for Great Whale River during the period 1701 to 1925 (Table 1). Catchpole indicates that the mean summer temperature of 1816 ($4.4^{\circ}C$)

was 1.34 standard deviations below the mean. Summer coldness in the following year was apparently even more severe. Such efforts to test the value of one type of proxy data (climatic records from Hudson's Bay Company documents) against another (tree-ring records) for the Hudson Bay - James Bay region is praiseworthy: in this case the data seem to corroborate each other. In his eclectic treatment of this critical period, Catchpole shows that Bay region ice conditions were indicative of prevailing northerly or northwesterly winds, pumping cold arctic air over the central and eastern parts of North America. But this is only one piece of the puzzle. I suggest that an international meeting on this critical period should be convened in order to round out the evidence - at least on a northern hemispheric scale. What were conditions like in Russia and western North America during that summer?

In her paper on the Little Ice Age on eastern Hudson Bay and James Bay, Cynthia Wilson again demonstrates the wealth of information in Hudson's Bay Company archives from 1814 to 1821. She stresses that the summers of 1816 and 1817 were not only colder than those on modern record, but were exceptionally severe even for that period. Like Catchpole, she mentions the unusually high frequency of northerly winds. Wilson gives some fascinating insights into the effects of the unrelenting sequence of cold summers from the autumn of 1815 to the spring of 1818: gardens failed; fish and game were scarce; adverse ice conditions threatened both the life-lines with England and internal distribution of food and trade; some sailors on English ships forced to overwinter died from scurvy and many natives died from starvation and cold; fuel needs at some posts approximately doubled; and Whale River was abandoned in 1816. The human impacts of such critical periods are worth revealing. Wilson's synoptic weather map for June 6, 1816 (Figure 7) provides an inkling of great things to come. Her second paper, "Daily Weather Maps for Canada, Summers 1816 to 1818 - a Pilot Study", providing a series of six daily weather maps for 1816 and summing up with one showing the paths of high and low pressure areas across central and eastern North America between June 3 and June 13, 1816 (Figure 6), is a remarkable achievement! To my knowledge, these are the first weather maps for Canada based on historic data and fulfill an earlier claim that the NMNS Climatic Change Project "is at a most interesting and important juncture" (Harington 1981). Wilson is presently augmenting the Hudson's Bay Company data with similar information from other sources in order to refine, and extend this series of 1816 weather maps.

Tim Ball's analysis of 1715-1802 weather data for Churchill and York Factory (e.g. number of days of rain, number of days with heavy or continuous rainfall, number of days of thunder, percentage frequencies of south and north winds) in Hudson's Bay Company documents point to

1760 as being a critical period in that region of Hudson Bay. Apparently, prior to 1760, the mean position of the Arctic Front was south of York Factory - placing both localities in the tundra zone - whereas, after 1760, Churchill remains in the tundra zone, while York Factory is now in the boreal zone. He states that 1760 could mark the end of the Little Ice Age in that part of North America, however there are no temperature records to indicate such an amelioration.

Baron and Gordon are members of an interdisciplinary team reconstructing New England climate for 1620-1980 using a variety of historical sources (e.g. early instrumental records, diaries, newspapers, agricultural journals, ships' logs and government documents). The reconstructions reveal that for eastern Massachusetts a general warming since the mid-1800s has been accompanied by an increased frequency of thunderstorms and snowfall, and a tendency to cloudier conditions. Future work will concentrate on other areas in New England and in statistical treatment of the data.

In the single paper on prehistory, Jim Wright notes that environment exerts an important influence on cultural systems that can be analyzed by archaeology (e.g. technology, subsistence and settlement pattern). Conversely, he argues that prehistoric cultural distributions can be treated as biological macrofossil evidence, so that cultural shifts can be interpreted as responses to environmental changes that may not be detected by fossil pollen records from a region. The potential of using shifting cultural patterns as a sensitive tool in detecting climatic change deserves further investigation.

Two papers deal with tree-ring evidence for changing climate. Gordon Jacoby and his coauthors use tree-ring data to reconstruct degree-days for central Alaska and northwestern Canada since 1574. The reconstruction corroborates subarctic glaciological data. A gradual warming trend since about 1840, with superposed decadal trends is detected for this part of northwestern North America.

Do nineteenth century tree-rings from 15 sites from Hudson Bay to the Pacific Coast respond to volcanic eruptions such as Tambora (1815) and Krakatau (1883)? Marion Parker shows that there is no obvious continent-wide reduction in ring-width and density for periods following those eruptions.

Seven papers, based on studies of fossil pollen, concern Late-Glacial and Holocene climate in eastern North America, Alberta and British Columbia. Joyce Macpherson synthesizes fossil pollen information for the Newfoundland and eastern Labrador-Ungava region, drawing climatic inferences from this and related evidence. Her study culminates in a series of maps

(Figure 6) showing suggested storm tracks for the early Holocene, 6000 yr B.P., and 5000-4000 yr B.P. Evidently optimal conditions existed from about 6000-3000 yr B.P., followed by climatic deterioration - probably related to increased dominance of dry arctic air masses.

Bob Mott detects evidence, based on fossil pollen studies in the Maritime Provinces, for warming of the region after deglaciation followed by climatic deterioration and a dramatic warming about 10,000 yr B.P. Evidence is strengthening that these oscillations are correlative with the Allerød warm interval, Younger Dryas cooling, and subsequent Holocene climatic amelioration of northwestern Europe.

Bartlein and Webb have undertaken a great challenge: to construct an isotherm map of mean July temperatures for eastern North America at 6000 yr B.P. Multiple regression techniques are used to obtain temperature estimates from pollen-indicated vegetational changes. At this critical period (part of the Hypsithermal for much of North America), the map shows that mean July temperatures were higher at 6000 yr B.P. than today for a region from central Québec to the southern United States. Temperature estimates are much lower than those of today in northern Québec - reflecting the reality of a small residual ice sheet in Ungava at 6000 yr B.P. Insofar as comparison of techniques is concerned, the pattern of temperature differences mapped is broadly similar to that of the National Center for Atmospheric Research global circulation model, which used July solar radiation values for 6000 yr B.P.

Three papers from Len Hills and coauthors deal with inferences about postglacial climate in southwestern Alberta and southeastern British Columbia based on fossil pollen evidence. At Waterton Lakes (Linnet Lake site), Alberta, evidently the period from 7000 to 5000 yr B.P. was warmer and drier than the present, 5000 to ?4000 yr B.P. cooler, ?4000 to 3000 yr B.P. warmer and drier, followed by a progressive cooling from 3000 yr B.P. to the present. At Crowsnest Pass (Crowsnest Lake site), Alberta, grasslands have occupied the area for about the last 10,500 years, increasing considerably in area during the Hypsithermal (shortly after 8000 yr B.P. to between 6000 and 5000 yr B.P.). Probably grazing conditions for ungulates (e.g. bison) and opportunities for human hunters were particularly favourable then. Cool, wet conditions followed, except for an interval of grassland expansion. Sediments from a bog (Bog A) in the Elk Valley of British Columbia have yielded a pollen profile extending back at least 13,500 years. Shrub-herb tundra seems to have been characteristic of the region following deglaciation. If sedimentation rates were steady, the Hypsithermal, (pollen Zones III and IV) occurred between about 9700-8700 yr B.P. and 8300-5700 yr B.P. Cooler, moister conditions followed (Zone V), as interpreted from a decrease in the pine to spruce plus fir ratio.

Apparently the heart of the Hypsithermal in this region of Alberta and British Columbia extended from about 7000 to 5000 yr B.P.

Rolf Mathewes concludes that the Hypsithermal in coastal and interior British Columbia occurred from approximately 10,000 to 7000 yr B.P. This is in contrast to the roughly 7000 to 5000 yr B.P. span for that critical period in southern Alberta mentioned by Hills *et al.*, and the 6000 to 3000 yr B.P. period mentioned by Macpherson for Newfoundland. Presumably such increasingly later dates for the Hypsithermal from western to eastern Canada do not constitute a simple, linear geographic pattern, but one that may be better explained in more complex terms - for example, by: (a) lack of a strict definition of "Hypsithermal"; (b) a need to consider the lag between climatic warming and drying and the actual regional palynological expression of the Hypsithermal, which may involve differing migration rates of various plant species combined with the differential effect of dispersal barriers.

Two papers deal with glacial geology and geochronology. The first, by John Andrews, presents a fascinating series of maps of the Baffin Island-Ungava area of the eastern Canadian Arctic, portraying changing environmental conditions over the past 11,000 years. Each map contains a great deal of paleoenvironmental information (e.g. localities with their radiocarbon dates, July temperature estimates based on pollen data, sea-ice and iceberg conditions, and the situation of warm and cold currents) - they are the next best things to a series of ancient air photographs! Andrews points out that although the history of glacial advances and retreats is rather well known for the last 8000 years, we have much to learn about earlier events.

The second paper, by John Matthews and Charles Schweger, notes that the widespread Old Crow tephra (volcanic ash) constitutes a "critical period" because it evidently marks an "instant" in time some 60,000 years ago over much of the unglaciated area of Alaska and the Yukon Territory. Glacial-geological (e.g. permafrost degradation) and paleobiological (e.g. fossil pollen and animals) all indicate a single pulse of warm climate about that time. Study of the Old Crow tephra has the potential for revealing synoptic climate for a time during the Wisconsin glaciation that is poorly known, as well as providing fresh insight to climatic events predating and postdating it.

Of two papers in the Paleoclimatology and Glaciology section, Bea Alt's focuses on the climate of the Queen Elizabeth Islands during the Little Ice Age. She notes that 1964 synoptic conditions resemble those hypothesized for 1550-1620 in the region, and she uses them as a key to unlock the summer long-wave pattern during 1550-1620 - characterized by weak ridges in the

Mackenzie-Keewatin area, over eastern Greenland and the Barents Sea, and by strong troughs down western Greenland, between Greenland and Spitzbergen to the British Isles, and in the East Siberian Sea (Figure 4). The author notes that it would be worth checking this synoptic climate recreation for the early part of the Little Ice Age in the Queen Elizabeth Islands against similar schemes derived from other ice-core parameters (e.g. electrolytic conductivity and pollen). Again, corroboration by different techniques is important in determining the value of preliminary paleoclimatic reconstructions.

In the final paper in this section (with one of the most stimulating titles I have seen!), Claus Hammer states how climate during the Wisconsin glaciation differed between the northern and southern hemispheres. Further, he speculates that crossing the threshold into the last glaciation, or a future one, may need only a small triggering effect - possibly volcanic activity.

My impression after reading these papers is that Quaternary paleoclimatology in northern North America is a healthy and growing science. Symptomatic of this condition are increasing efforts to: (1) check results of climatic reconstructions using one type of proxy data against others in order to strengthen the basic framework of this body of knowledge; and (2) convert proxy data derived from paleobiological studies to climatic parameters (e.g. summer temperature), thus allowing the data to be easily mapped and enabling us to discern more readily past climatic conditions and trends.

In conclusion, I wish to mention that the National Museum of Natural Sciences, stemming from activities of its Climatic Change Project, is planning a travelling exhibit "Canada's Changing Climate". The display will consist of three main parts: present, past and future climate. It is scheduled for completion in the spring of 1985. After its premiere at the National Museum of National Sciences in Ottawa next summer, it will begin a five-year nationwide tour.

REFERENCES

Parker, M.L., L.A. Jozsa, S.G. Johnson, and P.A. Bramhall. 1981. Dendrochronological studies on the coasts of James Bay and Hudson Bay. In: Climatic Change in Canada 2. Edited by: C.R. Harington. Syllogeus No. 33:129-188.

Harington, C.R. 1981. Introduction. In: Climatic Change in Canada 2. Edited by C.R. Harington. Syllogeus No. 33:5-9.

Mass, C., and S.H. Schneider. 1977. Statistical evidence on the influence of sunspots and volcanic dust on long-term temperature records. Journal of Atmospheric Science 34:1995-2004.

Pollack, J.B., O.B. Toon, C. Sagan, A. Summers, B. Baldwin, and W. Van Camp. 1976. Volcanic eruptions and climatic change: a theoretical assessment. Journal of Geophysical Research 81:1071-1083.

Taylor, B.L. 1978. Volcanic eruptions and long-term temperature records: an empirical search for cause and effect. M.Sc. thesis, University of Toronto. (A paper with the same title was published in Quarterly Journal of the Royal Meteorological Society 106:175-199, 1980.)

Wexler, H. 1951. On the effects of volcanic dust on insolation and weather. Bulletin of the American Meteorological Society 32:48-51.

AVANT-PROPOS

C.R. Harington[1]

Les participants réguliers et associés au programme d'étude des changements climatiques du Musée national des sciences naturelles se sont réunis trois fois depuis la création de ce programme en 1977. La réunion qui a eu lieu les 19 et 20 mai 1983 à l'Edifice commémoratif Victoria, à Ottawa, était notre quatrième et la première d'envergure internationale. Elle avait pour but de clarifier, à l'aide de diverses disciplines, la nature, les causes et les effets de certaines périodes cruciales de l'histoire climatique du quaternaire dans le nord de l'Amérique du Nord (par exemple la déglaciation, il y a environ 10 000 ans, l'hypsithermal, le petit optimum climatique qui a eu lieu aux environs de l'an 1000 de notre ère, la petite époque glaciaire qui a eu lieu entre 1500 et 1850 environ, les périodes d'importantes éruptions volcaniques et les années de froid intense, de sécheresse, etc.).

Fait digne de mention, Ridgeley Williams a présenté une plaque, au début de la réunion, à Morley K. Thomas, au nom du programme d'étude des changements climatiques du MNSN, pour son exceptionelle contribution à l'étude des changements climatiques au Canada.

Morley s'est initié à la météorologie alors qu'il fréquentait l'Université Western Ontario. Il a servi quatre ans comme prévisionniste au sein de l'Aviation royale du Canada. Après la fin de la Deuxième Guerre mondiale en 1945, il étudie la climatologie à Toronto où il obtient une maîtrise en 1949. Il fait ensuite carrière dans la Fonction publique fédérale. D'abord directeur des Opérations à la Division du climat, il devient en 1971 chef de la Direction des applications de la météorologie, passe ensuite au poste de directeur général des services centraux en 1976, avant de recevoir finalement une affectation spéciale, en 1979-1980, comme directeur général du Centre climatique canadien, à Downsview, puis de prendre sa retraite. Il travaille actuellement à la rédaction d'un livre sur l'histoire de la météorologie au Canada.

Morley a publié de nombreux articles scientifiques et de vulgarisation, ainsi que des livres sur la climatologie. Il a participé à de nombreuses conférences internationales et régionales sur la climatologie, a présidé de 1978 à 1982 la Commission pour la climatologie et les applications de la météorologie et a contribué à l'élaboration du programme climatologique mondial et du programme climatique canadien. Il s'intéresse surtout sur le plan professionel aux applications de la météorologie et aux changements climatiques. Dans cette optique, il est facile de comprendre le caractère vigoureux et pratique du message qu'il a transmis dans l'exposé qu'il a donné après la présentation de la plaque: ... "les données paléoclimatiques vont devenir de plus en plus importantes pour les climatologues et donc pour toute la société, comme source de connaissances à la fois théoriques et pratiques. Ces données constituent un élément vraiment essentiel à notre connaissance des climats passés et futurs. Je sais que votre travail est précieux et important et vous le savez aussi, mais il nous faut en convaincre un large auditoire. C'est *vous* qui êtes les mieux placés pour "vendre" votre travail, pour en démontrer toute la valeur".

---

[1] Division de la paléobiologie, Musée national des sciences naturelles, Musées nationaux du Canada, Ottawa, Ontario K1A 0M8

Ce volume sur le symposium est divisé de façon quelque peu arbitraire en différents chapitres dont le titre indique la ou les principales disciplines en cause: bibliographies; données fournies par les instruments; documents d'archives; préhistoire; anneaux de croissance des arbres; pollens fossiles; géologie glaciaire et géochronologie; et paléoclimatologie et glaciologie. La plupart des communications se fondent sur des analyses (sept) de pollens fossiles des périodes tardiglaciaire et holocène et sur des études de documents historiques, tirés notamment des archives de la Compagnie de la Baie d'Hudson. En ce qui concerne les époques étudiées, près de la moitié des communications portent sur la période holocène (et plus particulièrement sur l'hypsithermal) alors qu'environ un quart traitent du XIX^{e} siècle, plusieurs d'entre elles abordent l'année sans été (1816). Sur le plan géographique, on traite de toutes les régions du Canada, notamment la baie d'Hudson et l'Arctique, et on mentionne aussi la Nouvelle-Angleterre, l'Alaska et le Groenland.

Les bibliographies sont d'utiles instruments pour les chercheurs qui s'efforcent de recueillir des données de base sur les périodes cruciales de l'histoire climatique du quaternaire dans le nord de l'Amérique du Nord. Ce volume contient deux communications sur le sujet. La première, oeuvre de Martha Andrews, décrit le processus de sélection, de classement et d'informatisation des données utilisées pour l'*Annotated Bibliography of Holocene Paleoclimates*. Elle fait remarquer que le travail en est encore à l'étape préliminaire. La deuxième communication, rédigée par C.R. Harington, donne des précisions sur l'*Annotated Bibliography of Quaternary Climatic Change in Canada*. Sans doute peut-on observer des différences au niveau de la classification (par exemple, il y a 16 catégories dans le premier document comparativement à près de 40 dans le second, et près de deux fois plus de points de référence géographiques pour le Canada dans le premier que dans le second), mais il n'en reste pas moins que certaines conclusions présentent des similitudes frappantes (par exemple le nombre relativement élevé de références portant sur la géologie glaciaire et les pollens fossiles, et la quantité étonnante de documents consacrés aux îles de l'Arctique canadien). Bien entendu, on doit tenir compte de l'envergure différente des deux bibliographies. La première porte sur les quelque 10 000 dernières années à l'échelle planétaire, tandis que la seconde englobe les deux derniers millions d'années, mais uniquement au Canada.

Le phénomène climatique qui a le plus influé sur la société canadienne pendant le dernier siècle a sans doute été la sécheresse des années 1930 dans les Prairies. Berry et Williams indiquent que de tels cas ne sont pas exceptionnels. En fait, une sécheresse aussi grave que celle de la pire année de la période déjà mentionnée est susceptible de se produire une fois tous les 20 ans en moyenne. Cela démontre clairement la nécessité d'avoir des plans d'urgence soigneusement préparés afin d'affronter plus facilement à l'avenir de telles périodes critiques.

Les grandes éruptions qui ont eu lieu depuis celle du Krakatau, en 1883, ont-elles eu un effet sur le climat canadien? Walter Skinner jette un éclairage très nécessaire sur la question. En utilisant la méthode d'analyse relativement exacte qui procède par recoupement des témoignages d'ordres divers sur une période en particulier, laquelle a été mise au point par Mass et Schneider en 1977, l'auteur indique qu'il y a eu des baisses annuelles marquées de la température au Canada (et à un moindre degré des précipitations) après une forte éruption. Il constate que la baisse de température dans l'Arctique était légèrement supérieure à celle du reste du pays, ce qui équivalait à une chute d'environ 1°C tant au cours de l'année de l'éruption que de l'année subséquente. De plus, on a constaté une baisse d'environ 13 mm du niveau des précipitations dans les Prairies au cours de la période de culture (de mai à septembre) après des éruptions survenues sous des latitudes moyennes. Il est utile de comparer ce document avec celui de Marion Parker, dans lequel l'auteur tente de déterminer si les baisses de température auxquelles on aurait pu s'attendre à la suite des éruptions du Tambora, en 1815, et du Krakatau ont laissé des traces sur les anneaux annuels des arbres. L'auteur, qui fait appel à la technique de la dendroclimatologie, ne constate pas de répercussions à l'échelle du continent, mais admet qu'il faudra mener d'autres études. Par ailleurs, les résultats obtenus par Skinner appuient les résultats théoriques de Pollack et autres (1976) et les conclusions empiriques de Mass et Schneider (1977) et de Taylor (1978), mais non l'hypothèse des noyaux de condensation formulée en 1951 par Wexler suivant laquelle les fortes éruptions volcaniques seraient suivies de précipitations accrues.

Moira Dunbar, qui a étudié les journaux des explorateurs et des pêcheurs de baleines, fait valoir que les glaces étaient plus compactes pendant la période d'exploration intensive des îles de l'Arctique canadien (vers 1818-1860) qu'au cours du présent siècle. Ce facteur devrait être pris sérieusement en considération par les historiens lorsqu'il s'agit d'évaluer le succès ou l'insuccès des missions lancées à la recherche des restes de l'expédition de Sir John Franklin.

Les archives de la Compagnie de la Baie d'Hudson sont une mine de renseignements précieux sur le climat dans le centre et le nord du Canada au cours des cent dernières années. Les communications d'Alan Catchpole, Cynthia Wilson et Timothy Ball indiquent à quel point cette information peut être utile lorsqu'il s'agit de décrire la nature de la température et du climat dans les régions déjà mentionnées, notamment au cours du XIX^{e} siècle.

Alan Catchpole étudie dans sa communication "l'année sans été" (1816), et fait ressortir combien sont nombreux les documents d'archives de la Compagnie de la Baie d'Hudson portant sur cette région et cette période: dix-neuf journaux quotidiens de postes, neuf registres contenant des relevés quotidiens de la température, quatre livres de bord de navires et même un journal tenu sur un sloop côtier. L'auteur fait une utilisation intéressante des connées fournies par les anneaux des arbres du lac Cri (Québec) (Parker et autres, 1981), dont il se sert pour calculer les températures moyenne estivales de la région de la rivière Great Whale au cours de la période 1701-1925 (tableau 1). Alan Catchpole indique aussi que la température moyenne pour l'été 1816 (4,4°C) correspondait à un écart-type de 1,34 de moins que la moyenne. L'été suivant fut encore plus froid, semble-t-il. Ces initiatives visant à évaluer la valeur d'un type de données de seconde main (les relevés de températures provenant des archives de la Compagnie de la Baie d'Hudson) en regard d'un autre type d'informations (les anneaux des arbres) de la région de la baie d'Hudson et de la baie James sont louables; dans ce cas, les données semblent se corroborer. Catchpole fait ressortir dans son traitement éclectique de cette période cruciale que l'état de la glace dans la région de la baie d'Hudson indiquait la présence de vents dominants du nord ou du nord-ouest, qui faisait souffler l'air froid de l'Arctique sur tout le centre et l'est de l'Amérique du Nord. Toutefois, il ne s'agit là que d'un aspect du puzzle. Je propose la tenue d'une conférence internationale sur cette période cruciale afin de finir de rassembler toutes ces données - du moins à l'échelle de l'hémisphère nord. Quelles étaient les conditions climatiques en Russie et dans l'ouest de l'Amérique du Nord au cours de cet été-là?

Dans sa communication sur la petite époque glaciaire du côté est de la baie d'Hudson et de la baie James, Cynthia Wilson nous donne une autre occasion de constater à quel point les archives de la Compagnie de la Baie d'Hudson de 1814 à 1821 constituent une précieuse mine de renseignements. Elle insiste sur le fait que les étés 1816 et 1817 ne furent pas seulement plus froids que les étés modernes, mais qu'ils furent particulièrement mauvais même pour cette période. Tout comme Catchpole, elle mentionne la fréquence exceptionnellement élevée des vents du nord. Wilson fait mieux comprendre, au moyen d'exemples très bien choisis, les conséquences des terribles étés froids de 1816 et 1817: les jardins ne produisirent pas, le poisson et le gibier se firent rares, les glaces mirent en danger aussi bien l'unique voie de communication avec l'Angleterre que la distribution de la nourriture et les activités commerciales à l'intérieur, certains marins des navires britanniques qui durent hiverner sur place moururent du scorbut et de nombreux autochtones moururent de faim et de froid, la consommation de bois doubla presque à certains postes, et le poste de la rivière Whale fut abandonné en 1816. Les répercussions qu'eurent ces périodes si critiques sur les humains méritent d'être mentionnées. La carte météorologique synoptique du 6 juin 1816 (figure 7) établie par Wilson donne un avant-goût des belles réalisations futures de l'auteur. La deuxième communication de Cynthia Wilson, *Daily Weather Maps for Canada, Summers 1816 to 1818 - a Pilot Study*, qui comprend une série de six cartes météorologiques quotidiennes pour 1816 ainsi qu'une autre carte montrant uniquement les déplacement des systèmes de haute et de basse pression dans le centre et l'est de l'Amérique du Nord du 3 au 13 juin 1816 (figure 6), est une remarquable réussite. A ma connaissance, il s'agit des premières cartes météorologiques pour le Canada établies d'après des données historiques et ces documents viennent confirmer concrètement l'assertion selon laquelle le programme d'étude des changements climatiques "est arrivé à un tournant très intéressant et très important" (Harington 1981). Wilson complète actuellement les données de la Compagnie de la Baie d'Hudson par des renseignements similaires provenant d'autres sources, afin d'épurer et d'augmenter la série de cartes météorologiques pour 1816.

Analysant les données météorologiques sur la période de 1715 à 1802 relatives à Churchill et à York Factory (par exemple nombre de jours de pluie, nombre de jours où il a plu beaucoup où sans interruption, nombre de jours où il y a eu un orage, fréquence en pourcentage des vents du sud et du nord) qui figurent dans les documents de la Compagnie de la Baie d'Hudson, Tim Ball indique que 1760 est une date cruciale dans la région de la baie d'Hudson. Apparemment, avant cette date, la position moyenne du front arctique se situait au sud de

York Factory, qui, avec Churchill, se trouvait dans la zone de la toundra, alors qu'après 1760, Churchill demeure dans cette zone toundra et que York Factory passe à la zone boréale. Il précise que 1760 pourrait marquer la fin de la petite époque glaciaire dans cette région de l'Amérique du Nord, mais on n'a pas de données sur la température indiquant qu'il y a eu une amélioration.

Baron et Gordon sont membres d'une équipe interdisciplinaire qui s'efforce de reconstituer le climat de la Nouvelle-Angleterre de 1620 à 1980 au moyen de diverses sources historiques (par exemple les premières données fournies par les instruments, les journaux, les quotidiens, les périodiques sur l'agriculture, les journaux de bord et les documents officiels). Les chercheurs ont pu déterminer que le réchauffement général qui prévaut dans l'est du Massachusetts depuis les années 1850 s'est accompagné d'une augmentation de la fréquence des orages et des chutes de neige et d'un ennuagement plus fréquent. D'autres recherches porteront surtout sur d'autres régions de Nouvelle-Angleterre et sur le traitement statistique des données.

Jim Wright présente la seule communication sur la préhistoire, et il note que l'environnement a une importante influence sur les systèmes culturels qui peuvent être analysés par l'archéologie (par exemple la technologie, la subsistance et la mode de peuplement). Et réciproquement, il faut valoir que les vestiges laissés dans un endroit donné par une culture préhistorique peuvent être traités comme s'il s'agissait de données biologiques fournies par des macrofossiles, de sorte que les modifications culturelles peuvent être interprétées comme des réponses aux changements du milieu qui ne sont pas nécessairement révélés par les pollens fossiles d'une région. La possibilité d'utiliser des modifications culturelles pour connaître plus précisément les changements climatiques mérite de plus amples recherches.

Deux communications portent sur les anneaux des arbres comme témoins de changements climatiques. Gordon Jacoby et ses collaborateurs se servent des informations fournies par ces anneaux pour connaître les degrés-jours du centre de l'Alaska et du nord-ouest du Canada depuis 1574. La reconstitution obtenue corrobore les données glaciologiques subarctiques. On constate une tendance graduelle au réchauffement depuis 1840 pour cette partie de l'Amérique du Nord, tendance qui s'est maintenue dans toutes les décennies depuis cette date.

Au XIXe siècle, les anneaux des arbres provenant de 15 endroits, depuis la baie d'Hudson jusqu'à la côte du Pacifique, reflètent-ils les éruptions volcaniques comme celles du Tambora (1815) et du Krakatau (1883)? Marion Parker fait voir qu'il n'y a aucunne réduction évidente du diamètre et de la densité des anneaux à l'échelle du continent pour les périodes suivant ces éruptions.

Sept communications basées sur des études de pollens fossiles traitent du climat des époques tardiglaciaire et holocène dans l'est de l'Amérique du Nord, en Alberta et en Colombie-Britannique. Joyce Macpherson synthétise les informations sur les pollens fossiles pour Terre-Neuve et l'est de la région du Labrador-Ungava et elle tire des conclusions sur le climat à partir de ces données et d'informations connexes. Son étude se termine par une série de cartes (figure 6) qui propose les trajectoires de tempêtes au début de la période holocène, qui date de 6 000 B.P., et entre 5 000 et 4 000 B.P. Manifestement, les conditions climatiques optimales existaient entre 6 000 et 3 000 B.P. environ, période qui fut suivie d'une détérioration du climat, phénomène sans doute rattaché à la prédominance accrue des masses d'air arctique sec.

En s'appuyant sur des études de pollens fossiles dans les Maritimes, Bob Mott conclut qu'il y a eu un réchauffement après la déglaciation, puis une détérioration du climat et un réchauffement spectaculaire, il y a environ 10 000 ans. Il semble de plus en plus évident que ces changements ont un rapport avec l'oscillation d'Allerød, le refroidissement du Dryas supérieur et le réchauffement subséquent du climat à l'époque holocène dans le nord-ouest de l'Europe.

Bartlein et Webb se sont attaqués à une tâche fort difficile: établir une carte isotherme des températures moyennes de juillet pour l'est de l'Amérique du Nord, il y a 6 000 ans. Les auteurs utilisent les techniques de régression multiple pour obtenir des températures estimatives à partir des changements dans la végétation indiqués par les pollens. La carte établie pour cette période cruciale (qui se situait au moment de l'hypsithermal pour la majeure partie de l'Amérique du Nord) indique que les températures moyennes de juillet étaient plus élevées il y a 6 000 ans qu'aujourd'hui pour la région s'étendant du centre du Québec au sud des Etats-Unis. Cette température hypothétique est beaucoup plus faible que celle d'aujourd'hui dans le nord du Québec, ce qui s'explique par la présence d'une petite nappe de glace résiduelle en Ungava il y a 6 000 ans. En ce qui concerne la comparaison des techniques utilisées, le modèle des différences de température cartographiées est similaire en gros au modèle de circulation globale du Centre national de la recherche atmosphérique qui utilisait les valeurs du rayonnement solaire pour juillet de l'an 6000 B.P.

Trois communications par Len Hills et des collaborateurs formulent des hypothèses sur le climat postglaciaire dans le sud-ouest de l'Alberta et le sud-est de la Colombie-Britannique après examen de pollens fossiles. La température aux lacs Waterton (site du lac Linnet), en Alberta, pendant la période allant de 7000 à 5000 B.P. était manifestement plus chaude et plus sèche qu'actuellement, plus froide de 5000 à 4000 B.P. environ, plus chaude et plus sèche, de 4000 à 3000 B.P. environ, puis progressivement plus froide à partir de 3000 B.P. jusqu'à aujourd'hui. Les herbages occupent la région du col du Nid-de-Corbeau (site du lac Crowsnest) en Alberta depuis environ 10 500 ans et se sont accrus considérablement au cours de l'hypsithermal (période commençant peu après 8000 B.P. et s'étendant jusqu'entre 6000 et 5000 B.P.). Les conditions étaient alors sans doute particulièrement favorables aux ongulés (par exemple les bisons) qui pouvaient y paître et aux humains qui pouvaient y chasser. La température devint par la suite froide et humide, sauf pendant une courte période où les herbages gagnèrent du terrain. Des sédiments provenant d'une tourbière (tourbière A) située dans la vallée de la rivière Elk, en Colombie-Britanique, ont permis de découvrir un profil pollinique remontant à au moins 13 500 ans. Après la déglaciation, la région semble avoir été une zone de toundra caractérisée par des arbustes et des herbes. Si la sédimentation s'est effectuée à un rythme régulier, l'hypsithermal (zones polliniques III et IV) est survenu entre environ 9700-8700 B.P. et 8300-5700 B.P. La température est devenue par la suite plus froide et plus humide (zone V), comme le laisse supposer la diminution de la proportion des pins par rapport aux sapins et aux épinettes. Il semblerait que l'époque centrale de l'hypsithermal dans ces régions de l'Alberta et de la Colombie-Britannique se soit située entre environ 7000 et 5000 B.P.

Rolf Mathewes conclut que l'hypsithermal sur les côtes et à l'intérieur de la Colombie-Britannique s'est produit de 10 000 à 7000 B.P. environ. Cette datation contraste avec celle proposée par Hills et autres pour cette période cruciale dans le sud de l'Alberta, soit de 7000 à 5000 B.P. environ, et avec la période de 6000 à 3000 B.P. proposée par Macpherson pour Terre-Neuve. Le recul sans cesse croissant de la datation pour l'hypsithermal de l'ouest à l'est du Canada ne constitue vraisemblablement pas un modèle géographique simple et linéaire, mais une tendance qui peut être expliquée de façon plus complexe, par exemple grâce aux facteurs suivants: (a) l'absence d'une définition précise de l'hypsithermal et (b) la nécessité de considérer le décalage entre le réchauffement et l'assèchement du climat et la manifestation réelle de l'hypsithermal sur le plan palynologique dans une région donnée, ce qui peut comporter différents rythmes de migration pour diverses plantes et laisse entrevoir le rôle joué à cet égard par les obstacles physiques.

Deux communications traitent de la géologie glaciaire et de la géochronologie. La première, qui est de John Andrews, présente une très intéressante série de cartes de la région de la Terre de Baffin et de l'Ungava, dans l'est de l'Arctique canadien, et illustre les changements du milieu naturel au cours des 11 000 dernières années. Chaque carte contient un grand nombre de données paléoenvironnementales (par exemple la datation au carbone 14 de certains endroits, une évaluation de la température de juillet d'après les données fournies par les pollens, l'état de la glace de mer et des icebergs et la situation des courants chauds et froids) - c'est ce qu'il y a de mieux à défaut d'une série de photographies aériennes anciennes! Andrews souligne que même si l'histoire de la progression et du recul des glaciers au cours des 8000 dernières années est assez bien connue, il n'en reste pas moins qu'il nous en reste beaucoup à apprendre sur l'histoire antérieure.

Les auteurs de la deuxième communication, John Matthews et Charles Schweger, font remarquer que la tephra (cendre volcanique) des bords de la rivière Old Crow, qui est très répandue, témoigne d'une période cruciale parce qu'elle marque manifestement un instant très précis, il y a quelque 60 000 ans, où la majeure partie des régions de l'Alaska et du territoire du Yukon n'étaient pas recouvertes de glaciers. Toutes les données ayant trait à la géologie glaciaire (par exemple la dégradation du pergélisol) et de nature paléobiologique (par exemple les pollens et les animaux fossiles) indiquent une seule poussée de chaleur vers cette époque. L'étude de la tephra de la vallée de la Old Crow pourrait révéler un climat synoptique pendant une période de la glaciation du Wisconsin qui est mal connue et jeter un nouvel éclairage sur les changements climatiques antérieurs et postérieurs.

On trouve dans la section de la paléoclimatologie deux communications dont l'une, due à Bea Alt, se concentre sur le climat des îles de la Reine-Elizabeth pendant la petite époque glaciaire. L'auteur note que les conditions synoptiques de 1964 ressemblent à celles imaginées pour la période 1550-1620 dans la région et elle se sert de ces données pour tenter de reconstituer le modèle des grandes ondes d'été de 1550 à 1620, caractérisé par de faibles crêtes dans la région du Mackenzie-Keewatin et au-dessus de l'est du Groenland et de la mer de Barents, et par de fortes dépressions à partir de l'ouest du Groenland, entre cette île et le Spitzberg jusqu'aux îles Britanniques, et dans la mer de Sibérie orientale (figure 4). L'auteur note qu'il serait utile de comparer cette reconstitution du climat synoptique pour le début de la petite époque glaciaire dans les îles de la Reine-Elizabeth avec d'autres reconstitutions similaires établies à partir d'autres paramètres du noyau de la glace (par exemple la conductivité électrolytique et les pollens). Encore une fois, il est important d'utiliser différentes techniques afin de faire corroborer les reconstitutions paléoclimatiques préliminaires.

Claus Hammer est l'auteur de la dernière communication de cette section (qui a l'un des titres le plus stimulants que j'ai jamais vus!). Il y précise comment le climat qui régnait au cours de la glaciation du Wisconsin était différent dans les hémisphères nord et sud. De plus, il émet l'hypothèse qu'un simple phénomène, comme peut-être l'activité volcanique, aurait pu suffire à provoquer la dernière glaciation ou pourrait en entraîner une autre.

Après voir lu ces communications, j'ai l'impression que la paléoclimatologie du quaternaire dans le nord de l'Amerique du Nord est une science qui se porte bien et qui va de l'avant, à preuve les efforts croissants déployés pour: (1) vérifier la justesse des reconstitutions climatiques en comparant un type de données de seconde main à d'autres afin de renforcer les éléments de base de cet ensemble de connaissances; et (2) transformer les données de seconde main provenant d'études paléobiologiques en paramètres climatiques (par exemple les températures d'été). Ce faisant, on peut aisément cartographier les données et déterminer plus facilement les conditions et l'évolution du climat par le passé.

Finalement, j'aimerais mentionner que le Musée national des sciences naturelles est en train de préparer une exposition itinérante, *Les changements climatiques au Canada*, dans le cadre de son programme d'étude des changements climatiques. Cette exposition, qui comprendra trois parties principales, le climat passé, actuel et futur, sera vraisemblablement prête au printemps 1985. Après son inauguration, l'été prochain, au Musée national des sciences naturelles, à Ottawa, elle entreprendra une tournée de cinq ans qui l'amènera dans tout le pays.

REFERENCES

Parker, M.L., L.A. Jozsa, S.G. Johnson, et P.A. Bramhall. 1981. Dendrochronological studies on the coasts of James Bay and Hudson Bay. In: Climatic Change in Canada 2. Rédacteur: C.R. Harington. Syllogeus No. 33:129-188.

Harington, C.R. 1981. Introduction. In: Climatic Change in Canada 2. Rédacteur: C.R. Harington. Syllogeus No. 33:5-9.

Mass, C., et S.H. Schneider. 1977. Statistical evidence based on the influence of sunspots and volcanic dust on long-term temperature records. Journal of Atmospheric Science 34:1995-2004.

Pollack, J.B., O.B. Toon, C. Sagan, A. Summers, B. Baldwin, et W. Van Camp. 1976. Volcanic eruptions and climatic change: a theoretical assessment. Journal of Geophysical Research 81:1071-1083.

Taylor, B.L. 1978. Eruptions and long-term temperature records: an empirical search for cause and effect. Thèse de maîtrise en sciences, Université de Toronto. (Une communication portant le même titre a été publiée dans le volume 106 du Quarterly Journal of the Royal Meteorological Society 106:175-199, 1980.)

Wexler, H. 1951. On the effects of volcanic dust on isolation and weather. Bulletin of the American Meteorological Society 32:48-51.

M.K. Thomas

SPECIAL LECTURE

THE VALUE OF PALEOCLIMATIC DATA TO A CLIMATOLOGIST[1]

M.K. Thomas[2]

INTRODUCTION

Those of us who have been active in developing the Canadian Climate Program have followed the National Museum of Natural Sciences Climatic Change Project with much interest. We have read the collections of papers published in the *Syllogeus* series and appreciate the existence of such a substantial contribution to our climatic change literature. I congratulate Dr. Harington and associates in the National Museum for the development of the program and the persistence with which it has been pressed forward, despite resource restraints. I am happy to be here today and I anticipate learning much from the speakers and from the discussions.

I am here as a climatologist who began his career as a meteorologist and who over the years became more and more an administrator, at the expense of his science. My few words this morning will not fit into the specific topic of these meetings, but I do hope that I will get you thinking about the value of the work you are doing and convince you of the necessity of promoting and advertising your work and its results, more than you have ever done in the past.

Many of us, not directly in the paleoclimate sector, have been slow to recognize the value of paleoclimatic data. This is both my fault and your fault. Most scientists - probably including several of you - are woefully inept at promoting the value of their work, despite their marked devotion to science. This, of course, is traditional - most scientists are just not interested in business, public relations or promotion. But these days, when government budgets are on a stringent plateau, those who can document their programs, who can relate the value of their research findings to the needs of society and who can convincingly present their requests for resources, have a much better chance of launching and expanding programs than those who can't or won't. The scientist who buries himself in his science with the belief that benevolent management will look after him and his program will undoubtedly suffer. So recognize the value of your plans and programs, prepare convincing documents and speak up loudly and clearly at every opportunity.

---

[1] Ed. note: This address followed the presentation of a plaque, on behalf of the NMNS Climatic Change Project, to Morley Thomas in recognition of his valuable contributions to the study of climatic change in Canada. The lecture was given on May 19, 1983 in Ottawa.

[2] Canadian Climate Centre, Atmospheric Environment Service, Downsview, Ontario M3H 5T4

PALEOCLIMATIC DATA

The Atmospheric Environment Service Climate Data Archive has been accumulating in Toronto for over 100 years. Millions of bits of climatic data are stored there from thousands of official observing stations manned by professional and cooperative observers. There are, however, practically no paleoclimatic data in those archives. That this situation exists, is indicative of two facts: (a) climatologists, who have come into the field from meteorology, have not recognized data other than instrumental data from regular observing stations; and (b) scientists in tree-ring, ice core, pollen deposit, and other such studies leading to the collection of paleoclimatic data, have been too content to work in their own special fields without demanding a comprehensive approach to the cataloguing and archiving of *all* climatic data.

In defense of climatologists, I can report that most begin their professional careers as operational meteorologists working with current or real-time data transmitted from hundreds of observing stations as frequently as every hour. Only after moving into climatology do these meteorologists appreciate the existence of standard instrumental data archives, and they continue to be unaware of the existence of - let alone appreciate the value of - paleoclimatic data. On the other hand, scientists involved in pollen analysis, dendrochronology, etc., are most often true research scientists primarily interested in perfecting methods and in interpreting their experimental work into dated surrogate, substitute or proxy climatic data. The lack of understanding which still exists between climatologists and paleoclimatic scientists must be eliminated. Meetings such as this, and cooperation in programs such as the Canadian Climate Program, will help establish a truly national climate program. Paleoclimatic research and data must be recognized as an important part of the climate scene, your programs must be expanded and resourced and better relations forged with traditional climatologists.

THE VALUE

Paleoclimatic data are of great importance in that these data increase our knowledge of our present and past physical surroundings. Such knowledge is often considered intangible and so dismissed by some as not too important. To others, however, and I am one, climatic history is a most important aspect of all knowledge available to society. With increasing knowledge of climatic history we have a better understanding of all history - migrations, conquests, failures, successes and perhaps even an improved philosophy of life itself. Your contributions to this basic pool of knowledge are significant and should bring you great satisfaction. During some relatively prosperous and wealthy decades in the past, this would have been enough; senior bureaucrats would have honoured your requests for resources without much difficulty. In 1983, however, you have to make a stronger, more convincing case for the value of your paleoclimatic products if you hope to obtain resources to allow you to expand or even to carry on your current work.

Climatologists in the applied fields of agriculture, building design, transportation, water resources, etc. have a definite need for paleoclimatic data. Many applied climatologists, however, consider yours to be "proxy data" and nowhere near as valuable as instrumental data. Hence, in the past they have tended to ignore any paleoclimatic data which were available. However, instrumental data series are usually much too short to provide sufficient information regarding past extreme events, durations of drought, heat or cold and other climatic factors so important to designers and builders of capital projects expected to last safely for decades and centuries. Recent paleoclimatic data can be of great importance here.

Long thought unobtainable in our time by most responsible scientists, reliable climate forecasts for months, seasons, and years may be attainable within the next decade. Certainly our research climatologists have made great progress in their understanding and numerical modelling of the physical climate system. It is unlikely that recent paleoclimatic data will ever be available for sufficiently short time periods for direct use in numerical models designed for climatic prediction. However, much cooperation is needed between climate modellers and paleoclimatologists in determining and archiving information on surface temperature and surface moisture in the recent past. Such information will be used to test the "prediction" of past climates over large time scales of decades and centuries. Much work has already been done in the CLIMAP program on climates of 18,000 years ago and the work needs to be extended to reconstruct the climates of more recent times.

CONCLUSIONS

Mankind's knowledge of climatic history is today increasing at a phenomenal rate and credit for this must be given to you, the paleoclimatologist, not to the classical or applied climatologists nor to historians. In the future, paleoclimatic data will become increasingly important to climatologists and hence to society, both as classical and practical knowledge. The data are a most essential ingredient to our knowledge of past and future climates. I know that your work is valuable and important and so do you, but we have to convince a wide circle of people. *You* can be the most valuable salesman or promoter your work can ever have.

HOLOCENE PALEOCLIMATES: AN ANNOTATED BIBLIOGRAPHY

Martha Andrews[1]

INTRODUCTION

From the beginning of the National Museum of Natural Sciences Climatic Change in Canada Project, recognition has been given to the importance of bibliography to research; publication of the "Annotated Bibliography of Quaternary Climatic Change in Canada"[2] is eagerly awaited. The bibliography under discussion in this paper, sponsored by the United States Department of Energy (DOE) and being prepared at the University of Colorado, covers some of the same times and places that are being covered by the Climatic Change in Canada Project.

The intent of the original proposal for compiling an "Annotated Bibliography of Holocene Paleoclimates", to support ... the publication of an annotated bibliography of well-dated and well-interpreted proxy climatic records which would contribute greatly to our current knowledge about the spatial and temporal variations in Holocene climates" (Andrews and Andrews 1980), has been adhered to in principal all along, although some of the details have been altered to suit circumstances. The main objective of the project has been to provide the scientific community with a reference document which will be a starting point for research in an interdisciplinary field where publications are widely scattered and thus difficult to access.

THE INQUA PILOT STUDY

Prior to submission of the proposal to DOE, a pilot study was funded by the International Quaternary Association (INQUA) Palaeoclimate Commission which permitted investigation of the usefulness of the commercially available, computerized bibliographic databases for research concerning Holocene paleoclimates. Manually produced, printed, published indexes, with abstracts, were also evaluated.

---

[1] Department of Geological Sciences, University of Colorado, Campus Box 250, Boulder, Colorado 80309

[2] See the following paper in this volume.

The object of this pilot study was to find as many references as possible containing data relating to Holocene paleoclimate for a sample area, North Slope/Brooks Range, Alaska. Indexes and databases chosen for testing of their applicability were: "Arctic Bibliography", "CRREL Bibliography", Geo Abstracts, "Annotated Bibliography of Quaternary Shorelines", GEOREF, GEOARCHIVE, SCISEARCH and COLD REGIONS. These choices were based on experience of the investigators with similar studies, both first hand and from the literature. Papers published between 1960-1980 were sought.

Search strategy for the manual databases relied mainly on the availability of area-based indexes or classifications. Search strategy for the computerized databases consisted of (Brooks Range OR North Slope) AND (Holocene) AND (Glaci# or Climat#).

As a result of this search, a total of 73 unique references was produced, plus an additional 26 duplicates of 16 of the 73 items (Table 1). Of the 57 items found only once, 25 were found in "Arctic Bibliography", as were 13 of the 16 records duplicated. The next largest numbers found were 17 from Geo Abstracts, followed by nine from GEOREF (plus two of three duplicates not occurring in "Arctic Bibliography"). Thus, if only these three databases ("Arctic Bibliography", Geo Abstracts, and GEOREF) has been used, 66 (25+13+17+9+2) of the total 73 citations would have been found. "Annotated Bibliography of Quaternary Shorelines" was the source for four of the seven remaining items. All items except one from the "CRREL Bibliography" were picked up from the other databases.

Use of an area term in all search strategies could have biased results of the computer searches toward GEOREF - the only database with an areal index.

The results of this INQUA pilot study (Andrews 1980), which demonstrated that use of selected printed indexes and computerized databases could indeed produce a useful bibliography dealing with Holocene paleoclimates, were very valuable in writing the proposal for DOE, and served as a guide during the project at several points.

SELECTION OF MATERIALS FOR THE ANNOTATED BIBLIOGRAPHY ON HOLOCENE PALEOCLIMATES

Background

Papers reporting on research in polar and alpine areas, believed to be most sensitive to climatic change from a possible buildup of CO_2 in the atmosphere, have been chosen for the bibliography based on their inclusion of chronological and/or quantitative information. Approximately 1200 papers are being selected.

TABLE 1: *INTERNATIONAL QUATERNARY ASSOCIATION (INQUA) PILOT STUDY RESULTS*[1].

	Single Records	Duplicates	Singles & Duplicates	Distribution of the Duplicates[2]															
				1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16
Manual:																			
Arctic Bibliography	25	13	38	x	x	x	x	x	x		x	x	x	x	x	x	x		
CRREL Bibliography	1	6	7		x			x			x	x		x			x		
GeoAbstracts	17	9	26	x	x		x		x		x	x	x		x			x	
Annotated Bibliography of Quat. Shorelines	4	1	5			x													
Computer:																			
GEOREF	9	8	17					x	x	x	x	x		x	x			x	
GEOARCHIVE		1	1							x									
SCISEARCH		2	2							x									x
COLD REGIONS	1	2	2													x			x
Totals	57	42	99	2	3	2	2	3	3	4	4	2	3	3	3	2	2	2	2

---

[1] Source: Andrews (1980).

[2] Numbers 1-16 indicate number assigned to each duplicate citation for reference.

It was decided to limit inclusion based on date of publication. Since techniques have improved so much during the past 20 years for detection of climatic changes using proxy records, 1960 became the cutoff date for publication.

Technology in the information sciences has also improved vastly during the past 20 years; however the main impact of the commercially available bibliographic databases was not felt until the early 1970s, with a few exceptions, and it was therefore necessary to search printed indexes in order to go backward in time to 1960. Another good reason for using the printed indexes is that some are area-specific, and others have good area indexes, a characteristic notably lacking in the computerized databases.

Printed Indexes

During the first year of the project, one of the main areas of focus was that of searching printed indexes for materials relevant to the aims of the bibliography (Appendix 1). Basic to all of the searching strategies was the need to merge three types of terms which were required to appear simultaneously in all papers chosen for inclusion in the bibliography: (1) time; (2) subject; and (3) area.

As determined from the INQUA study detailed above, the printed indexes likely to prove most useful seemed to be the "Arctic Bibliography" and Geo Abstracts, both of which reach back in time well beyond our chosen starting date of 1960.

The "Arctic Bibliography" was searched from volumes 10 through 16, covering publications from 1960 to 1972 when collecting for that bibliography ceased. Some 200 records representative of the arctic areas in all countries surrounding the north pole, written in many languages, and interdisciplinary in subject, were thus acquired, complete with abstracts written originally for the "Arctic Bibliography". Since the "Arctic Bibliography" is by nature area-specific, search terms indicating the desired time frame and subject fields were used.

Two sections of the Geo Abstracts bibliographic series were very useful sources. The first of these comprises Geomorphological Abstracts, for which the cumulative index for 1960-1965 was searched; Geographical Abstracts A, Geomorphology, for which the cumulative index for 1966-1970 was searched; succeeded by Geo Abstracts A, Landforms and the Quaternary, for which the 1971-1975 index was searched. These KWIC (augmented) indexes proved very satisfactory. The second section of Geo Abstracts used includes Geographical Abstracts B, Biogeography and Climatology, for which the 1966-1970 index was searched, and Geo Abstracts B, from 1970-1973 subtitled Biogeography and Climatology, and since 1974 subtitled Climatology and Hydrology, which was searched via its 1971-1975 cumulative index.

Materials were found under the headings indicative of time, such as: Holocene; Flandrian; late combined with Quaternary, Wisconsin, Cenozoic, Cainozoic, or glaci#; post combined with Pleistocene, Wisconsin, or glaci#; Recent; and Present. These time terms were sought in combination with subject terms such as climat# or glaci# or environment, paleo# or palaeo# combined with botan# or climat# or ecolog# or environment or geograph# or sol# or temperature#. In searching the Geo Abstracts bibliographies, search terms were confined to time and subject terms; items fitting those two categories were then checked for location.

The "Antarctic Bibliography" was searched from volume 1-11, covering 1962-1980. The search terms paleoclimate, paleobotany, and ecology were the most useful; some 55 items were collected from this source. This again being an area-specific bibliography, subject terms were used for searching, and verification as to the desired time-frame followed.

The other major source of materials was the private reprint collection of John T. Andrews, from which a few hundred papers were selected.

Three additional printed bibliographies were used, each being useful for several dozen items. These are: "Baffin Island Quaternary Environments: an Annotated Bibliography"; "An Annotated Bibliography of Quaternary Climatic Changes in Canada" (preliminary draft of 1979); and "Ecology of the Canadian Arctic Archipelago: Selected References", 1976-1981, volumes 1-9.

Computerized Databases

Introduction

While undertaking research for this project, I visited the Boreal Institute for Northern Studies (BINS), at the University of Alberta, Edmonton. Several searches were run, on their computer, using the SPIRES and QL information systems. Files Boreal and ASTIS, representing parts of the collections at BINS and The Arctic Institute of North America in Calgary, were searched while experimenting with various search strategies. Some records were obtained in this manner, but the experience was valuable mainly in determining the types of search strategies likely to prove most useful later in examining the very large databases.

Selection of Databases

The process of choosing which computerized databases to use was carried out in two stages. First, based on my experience, a dozen databases were chosen for "testing" against a database called DIALINDEX (Appendix 1). This database is an index to all of the other databases available through Dialog Information Retrieval Services, the vendor of all of the databases used for this project. Second, the 12 chosen databases were combined into three groups representing general scientific subjects: biological/agricultural, marine, and geophysical. For each group, time terms plus a somewhat different selection of subject terms, depending on the scientific discipline, were searched. By running the selected search terms through the selected databases, 10 databases were identified as being worthy of a full-scale search. These databases are:

AGRICOLA (Files 110, 10)
BIOSIS PREVIEWS (Files 55, 5)
COMMONWEALTH AGRICULTURAL BUREAUX ABSTRACTS (File 50)
COMPREHENSIVE DISSERTATION INDEX (File 35)
CONFERENCE PAPERS INDEX (File 77)
GEOARCHIVE (File 58)
GEOREF (File 89)
NTIS (File 6)
OCEANIC ABSTRACTS (File 28)
SCISEARCH (Files 34, 94, 156, 187, 188)

Search Strategies

The results of the DIALINDEX search, in addition to being helpful in database selection, were also helpful in formulating the database search strategies, since the number of postings (or times the term appeared in the database) for each search term in each database searched on DIALINDEX is given. The terms showing the most postings were used further; those with few postings were eliminated.

Unfortunately, it was not possible with the computerized searches to include an area-search term, since these databases are not indexed by area in a reliable fashion. Even GEOREF, a partial exception to this limitation, is indexed to area using such specific terms that even the use of a very large number of area terms would not have given adequate returns.

Table 2 shows the search as it was run. Five different search strategies were used on the several databases, which were combined into five groups: biological/agricultural, oceanographic, general category databases, and geological (two).

The searches are basically constructed to yield, in the end, a number of retrieved items which combine any of the time terms with any of the subject terms, using the Boolean operators OR and AND.

The "winners", by a large margin, were GEOREF, GEOARCHIVE, and BIOSIS PREVIEWS. Table 3 shows the search results obtained for the biological and geological sciences.

BIOSIS PREVIEWS is covered by two files, File 55 covering publications dated 1969-1976, and File 5 covering 1976 to present. Sets 1 and 2 are the time terms searched, and sets 3-10 represent the subject terms. Set 11 combines the time and subject terms, and on File 5 gives a total of 214 items indexed to any combination of the time and subject terms searched. File 55 results show 64 items retrieved.

Geological databases searched were GEOARCHIVE (1974-1982) and GEOREF (1961-1982), which had shown such a high rate of response during the DIALINDEX search that they were searched in a deliberately narrow fashion.

GEOARCHIVE was searched simply using Holocene and Paleoclimatolog?, resulting in 374 retrievals. The GEOREF search was narrowed even further, using Holocene and Paleoclimatolog?/DE*. The DE* indicates that retrievals should be limited to papers in which Paleoclimatolog? is the main topic of the paper. Even with such limitations, 1225 items were retrieved from GEOREF.

Up to this point, materials have been used from BIOSIS PREVIEWS, OCEANIC ABSTRACTS, and NTIS. Only a few dozen items have been used from these files, with varying levels of satisfaction to the author! It is expected that a few hundred more will be used from the geological databases, but nowhere near the nearly 1600 "found" by the computer searches.

The basic reason for the overload in retrieval is the inability of any of the computerized databases to be searched by area terms successfully. Even GEOREF has too many pitfalls to be reliable. Also, the computer search produced many duplicates of materials already found in the printed indexes and many papers in foreign languages without English abstracts or indication of the existence of an English translation. The lack of abstracts in GEOREF and GEOARCHIVE reduces the usefulness of their records to simply an indication of where a paper is to be found. These leads are being followed for further collection of what is estimated to be a few hundred papers.

TABLE 2: *SEARCH STRATEGY FOR COMPUTERIZED BIBLIOGRAPHIC DATABASES.*

```
Select files 10, 110, 5, 55, 50

SS (HOLOCENE OR LATE(W)QUATERNARY) AND (PALEOECOLOG? OR PALYNOLOG? OR RADIOCARBON(W)DAT? OR
   DENDROCHRONOLOG? OR PALEOBOTAN? OR AGRICULTUR? OR ARCHEOLOG? OR ARCHAEOLOG?)
```

```
Select file 28

SS (HOLOCENE OR LATE(W)QUATERNARY) AND (PALEOCLIMATOLOG? OR PALEOOCEANOGRAPH? OR SEA(W)LEVEL?
   OR DEEP(W)SEA(W)SEDIMENT?)
```

```
Select files 35, 77, 6, 34, 94, 186, 187, 188, 65

SS (HOLOCENE OR LATE(W)QUATERNARY) AND (PALEOCLIMATOLOG? OR PALEOECOLOG? OR PALYNOLOG? OR
   PALEOTEMPERATURE? OR GEOCHRONOLOG? OR GLACI?(W)GEOLOG OR RADIOCARBON(W)DAT? OR PALEO-
   OCEANOGRAPH? OR SEA(W)LEVEL? OR DEEP(W)SEA(W)SEDIMENT? OR PALEOBOTAN? OR AGRICULTUR?)
```

```
Select file 89, GeoRef

SS HOLOCENE AND PALEOCLIMATOLOG?/DE*
```

```
Select file 58 GeoArchive

SS HOLOCENE AND PALEOCLIMATOLOG?
```

TABLE 3: *SEARCH RESULTS OBTAINED FOR EXAMPLES IN THE BIOLOGICAL AND GEOLOGICAL SCIENCES.*

BIOSIS PREVIEWS 1969-1976

```
User24539   Date:13dec82   Time:14:48:47   File: 55

Set Items Description
  1    391 HOLOCENE
  2     85 LATE(W)QUATERNARY
  3     14 PALEOECOLOG?
  4    805 PALYNOLOG?
  5     23 RADIOCARBON(W)DAT?
  6      4 DENDROCHRONOLOG?
  7    132 PALEOBOTAN?
  8   3973 AGRICULTUR?
  9     80 ARCHEOLOG?
 10    491 ARCHAEOLOG?
 11     64  (1 OR 2) AND (3 OR 4 OR 5 OR 6

Print 11/5/1-64

Search Time:  0.029   Prints:  64   Descs.:  12
```

BIOSIS PREVIEWS 1976-1983

```
User24539   Date:13dec82   Time:14:46:47   File:  5

Set Items Description
  1    677 HOLOCENE
  2    153 LATE(W)QUATERNARY
  3    250 PALEOECOLOG?
  4   1007 PALYNOLOG?
  5    227 RADIOCARBON(W)DAT?
  6     40 DENDROCHRONOLOG?
  7    165 PALEOBOTAN?
  8   5085 AGRICULTUR?
  9     99 ARCHEOLOG?
 10    505 ARCHAEOLOG?
 11    214  (1 OR 2) AND (3 OR 4 OR 5 OR 6

Print 11/5/1-214

Search Time:  0.054   Prints: 214   Descs.:  12
```

GEOARCHIVE 1974-1982

```
User24539   Date:13dec82   Time:15:38:36   File: 58

Set Items Description
  1   3838 HOLOCENE
  2   2529 PALEOCLIMATOLOG?
  3    374 1*2

Print 3/5/1-374

Search Time:  0.033   Prints: 374   Descs.:   2
```

GEOREF 1961-1982

```
User24539   Date:13dec82   Time:15:45:37   File: 89

Set Items Description
  1  14713 HOLOCENE
  2   7042 PALEOCLIMATOLOG?/DE*
  3   1225  1 AND 2

Print 3/5/1-1225/AU

Search Time:  0.115   Prints:1225   Descs.:   2
```

COMPUTERIZATION OF THE PROJECT

Very early in the project, in fact during the writing of the proposal, hope was expressed that the material selected and collected could be stored in machine-readable form, without a very clear idea of just how this would be accomplished. One reviewer of the proposal expressed the desire to see a "modern" method of storage and retrieval used.

Fortuitously, this hope became a real possibility, when, at the suggestion of the project technical monitor, negotiations were furthered between the author and the Director of the Carbon Dioxide Information Center (CDIC) at Oak Ridge National Laboratory. An informal agreement to collaborate has resulted in the addition of the "Annotated Bibliography of Holocene Paleoclimates" to the automated CDIC Bibliographic Information System (BIS) which already contained over 2000 citations covering many subject areas related to CO_2. I emphasize that a text database the size of the "Annotated Bibliography of Holocene Paleoclimates" can only be manipulated successfully on a mainframe computer. In addition to providing online access to this bibliography, cooperation with CDIC includes computer production and printing of the bibliography and creation of the many useful indexes described later in this paper.

Several options were considered for transfer of machine-readable information to the IBM 3033 computer at Oak Ridge. The method finally chosen was the one I favoured from the beginning, that of using a microcomputer to input text to diskettes to be sent by mail to CDIC and read by the identical microcomputer via a transfer program into the IBM 3033.

Selection of a microcomputer for this task was simple. Since CDIC already possessed a Radio Shack TRS 80 III, this was the model used. Using SUPERSCRIPSIT, a Radio Shack word-processing program, a prompt screen was set up and materials have been successfully entered.

FORMAT DECISIONS

Table 4 shows the draft worksheet used for "The Holocene Paleoclimate Bibliography" input.

The first three fields are for entry of numerical codes: INPUTEAM HOLOCENE is the project code name for CDIC purposes. The fields LIT TYPE, PUB DATE, AUTHOR, TITLE, F TITLE and LANGUAGE are readily recognizable as: type of document: report, journal, etc.; publication date; author or authors; title in English; title in foreign language if given; and language, if other than English (notation of a second language summary or abstract is also given here).

TABLE 4: *HOLOCENE PALEOCLIMATE BIBLIOGRAPHY INPUT.*

ADSEP #	
ABHP #	
JTA #	
INPUTEAM	HOLOCENE
LIT TYPE	Journal
PUB DATE	
AUTHOR	
TITLE	
F TITLE	
PUB DESC	
LANGUAGE	
SUBJ CAT	
GEOG DESC	
TIME	Holocene
DAT METH	
KEYWORDS	
ABSTRACT	
COMMENT	
SEC SOUR	

The other fields are of more interest, since it is in them that each document is described in terms giving "added value" to the original publication.

SUBJ CAT. Sixteen "methods of reconstruction" were defined, and it is by these terms that the bibliography will be arranged for publication. More than one term may be assigned to a bibliographic record. The primary one will be used for classified arrangement of the record, and all terms will also be used as indexing terms. The methods of reconstruction of paleoclimates used are as follows:

PALYNOLOGIC
PALEOBOTANIC
PALEOZOOLOGIC
PEDOLOGIC
AGRICULTURAL
ARCHEOLOGIC
HISTORIC
STRATIGRAPHIC
SEDIMENTOLOGIC
GLACIOLOGIC
GLACIAL GEOLOGIC
OCEANOGRAPHIC
CLIMATOLOGIC
DENDROCLIMATOLOGIC
GEOLOGIC
GEOMORPHOLOGIC

GEOGDESC. Much consideration went into determining the approximately 75 regional divisions used as geographical descriptors. Each area is delineated and shown on one of two maps to be published with the bibliography. These maps are polar projections of the Arctic and Antarctic regions at a scale of 1:17,000,000 with an inset to show alpine areas. It was decided to break the area down into regions mainly based on political boundaries (for ease of reference to larger scale maps), but including some breakdowns by physical areas such as island groups. A few of these were broken down further based on: (1) the supposition that reference to an area as large and diverse as Greenland would not be very useful; and (2) the fact that some areas, such as Alaska and Baffin Island, contain such diverse physical areas, and have so many references to them, that further breakdown would be useful. It was a compromise between trying not to put too many references into a single region, and not having regions so large that referencing them was pointless. Nevertheless, both the USSR and Antarctica were finally divided into only two regions each.

TIME. Each bibliographic record is described as to time within the Holocene, using time-stratigraphic terms. It was proposed to key each record to a table for the printed volume. However, this proved impractical and, instead, a table will be constructed graphing the frequency of appearance of each time term (one descriptor to a paper) in the bibliography. Climatic-event terms such as Neoglaciation and Little Ice Age are used as keywords.

DAT METH. Methods used by the researchers to provide dates for paleoclimate reconstructions are entered for each record; any number of these terms that are appropriate may be used, and will be used as indexing terms. These terms form a controlled vocabulary and include radiocarbon dating, correlations, etc.

KEYWORDS. The keywords are basically subject terms used to describe the document. A thesaurus devised at CDIC was used as much as possible, but many terms were added as needed. These terms will provide an index to the printed volume, and will be a main point of access for the online users.

ABSTRACT. Every record will have an abstract. Many of these abstracts will be written originally by John T. Andrews. Others will be author abstracts; some author abstracts have been modified. Additionally, many abstracts have been quoted from secondary sources. In all cases, credit is given to the source of the abstract.

COMMENT. In this section, notations are entered of anything extra thought to be valuable to the user.

SEC SOUR. Secondary sources used, such as the "Arctic Bibliography", are given in this field.

Table 5 shows an example of a finished worksheet, completely entered on the microcomputer.

TABLE 5: *WORKSHEET COMPLETED ON THE MICROCOMPUTER.*

```
<582>
<ABHP>60
<INPUTEAM>HOLOCENE
<LIT TYPE>Journal
<PUB DATE>1963
<AUTHOR>Mackay, J.R.; Terasmae, J.
<TITLE>Pollen Diagrams in the Mackenzie Delta Area, N.W.T.
<PUB DESC>Arctic 16(4):228-38
<SUBJ CAT>Palynologic; Geomorphologic
<GEOGDESC>Mackenzie District
<TIME>Early Holocene
<DAT METH>Radiocarbon dating
<KEYWORDS>POLLEN ANALYSIS; CLIMATIC CHANGES; PEAT
<ABSTRACT>Considers postglacial climatic changes in northern Mackenzie District, from
palynological and other evidence. Peat from exposures in two areas, also alluvial sediments
from a drill hole were analyzed; pollen diagrams with radiocarbon control are presented for
the peat. From these investigations and geomorphic interpretation of the exposures, fossil
evidence, etc., a tentative climatic sequence is proposed; deglaciation about 12,000 yrs.
ago, a cool-dry climate 8500-7500 B.P., a warmer and drier period, and increased moisture and
cooling climate in late postglacial time. The last changes are indicated by increases in
alder and Ericaceae and the formation of pingos. (AB80944)
<SEC SOUR>AB80944
<583>
<ABHP>649
<INPUTEAM>HOLOCENE
<LIT TYPE>Journal
<PUB DATE>1966
<AUTHOR>McCulloch, D.; Hopkins D.
<TITLE>Evidence for an Early Recent Warm Interval in Northwestern Alaska
<PUB DESC>Geological Society of America  Bulletin 77(10):1089-1107
<SUBJ CAT>Paleobotanic; Pedologic
<GEOGDESC>Alaska, North Slope
<TIME>Early Holocene
<DAT METH>Radiocarbon dating
<KEYWORDS>WARMING TRENDS; GLACIER ADVANCES; FOSSILS; PALEOSOLS; PERIGLACIAL FEATURES
<ABSTRACT>In the coastal tundra covered area of N.W. Alaska, a warm interval that began at
least 10,000 years B.P. and lasted until at least 8,300 years ago is recorded by the presence
of fossil wood, fossil beaver-gnawed wood found beyond the modern range of beaver, evidence
of ice-wedge melting, buried soils and soils that extend below the top of modern permafrost.
Dating of the warm interval is based on eight radiocarbon dates. Although these do not
provide tight control for either the beginning or the end, they permit the interpretation
that the warm event began at the start of the world-wide postglacial warming and ended at the
time of the Anvik Lake glacial readvance in the Brooks Range. If this is correct the early
Recent warm interval and Livingstone's postglacial thermal maximum in the Brooks Range were
separated by a period of cooler climate. It is suggested that the postglacial thermal maximum
between 6000 and 3000 B.P. recorded in the Brooks Range is poorly shown in the coastal areas
of Northwestern Alaska, because of the lower summer temperatures of accompanying maritime
climate.  (Auth)
<SEC SOUR>GA 67A/1481
```

ANALYSIS OF BIBLIOGRAPHIC RECORDS TO DATE

The main goals of this paper are not only to show how this bibliography is being constructed, but also to show the types of information that can be obtained through analysis of the groupings produced under various types of indexing terms.

Of a projected total of some 1200 entries, over one-third (465) have been processed by CDIC and printed out, with indexes to:

(1) all authors

(2) all subject categories

(3) all keywords

(4) geographic descriptors (mainly for the online user who will not be able to refer to the maps)

(5) time,

(6) dating method, and

(7) a permuted title index

The two maps already described, and the graph of time terms will appear only as part of the printed volume.

The groupings apparent in these indexes are of some interest. In Table 6, by looking at the number of times a certain subject category ("method of reconstruction") has been assigned to a record (more than one can be assigned to a record), an idea can be gained of which methods are being used most commonly in paleoclimatic reconstruction - in order these are Glacial geologic, Palynologic, Climatologic, Glaciologic, Paleobotanic, and Stratigraphic.

Table 7 lists the geographic descriptors used in the area index, with figures showing how many times each has been used. Rather than attempt an analysis of the circumpolar pattern, attention is focused on northern Canada (Figure 1). It may be seen that the Canadian Arctic Archipelago has been covered in the most concentrated manner, with 25 references to Ellesmere Island and the Queen Elizabeth Islands, and 44 to Baffin Island. The rest of Canada is evenly covered with 1-9 references to each area. For Canada as a whole, including bibliographic records indexed to broader areas such as "Canada", "Arctic", and "North America", a total of 158 papers has been entered, approximately one-third of the total of 465 papers for which indexes are available at this point. However, it is to be remembered that this number represents only one-third of the projected total.

TABLE 6: *SUBJECT CATEGORIES.*

(Number beside term indicates number of times the term has been assigned; more than one may be assigned to a paper. Numbers 1-6 in parentheses indicate rank order of six most frequently used terms.)

Agricultural	2
Archeologic	20
Climatologic	71 (3)
Dendroclimatologic	14
Geologic	7
Geomorphologic	26
Glacial geologic	161 (1)
Glaciologic	58 (4)
Historic	35
Oceanographic	10
Paleobotanic	42 (5)
Paleozoologic	16
Palynologic	86 (2)
Pedologic	8
Sedimentologic	2
Stratigraphic	38 (6)

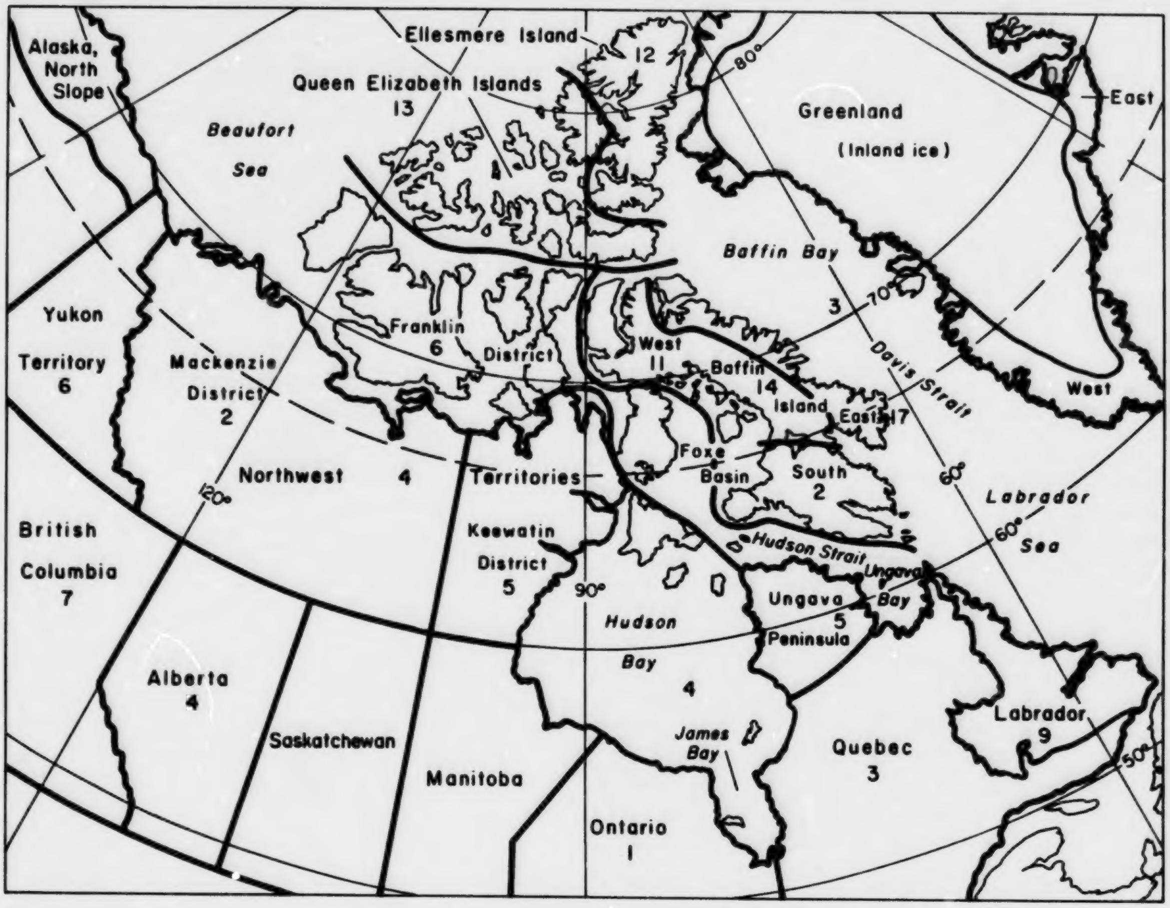

Alaska, North Slope
Ellesmere Island
12
80°
Queen Elizabeth Islands
13
Beaufort Sea
Greenland
(Inland ice)
East
Baffin Bay
3
70°
Yukon Territory
6
Mackenzie District
2
Franklin
6
District
West
11
Baffin
14
Island
East
17
Davis Strait
West
Foxe Basin
South
2
60°
Labrador Sea
Northwest
4
Territories
120°
British Columbia
7
Keewatin District
5
90°
Hudson Strait
60°
Ungava Bay
Ungava
5
Peninsula
Hudson Bay
4
Alberta
4
Saskatchewan
Manitoba
James Bay
Quebec
3
Labrador
9
50°
Ontario
1

FIGURE 1: *Map corresponding to the index to geographic descriptors (Table 7), showing Canadian regional descriptors and numbers of references to each region in the "Annotated Bibliography of Holocene Paleoclimates".*

Time terms are assigned only once to a paper. Of 465 papers, the time breakdown was as follows: Holocene (the entire Holocene)--235; Present time (twentieth century) 53; A.D. (including the Little Ice Age) 68; and the Sub-Atlantic through Pre-Boreal periods, two each in three divisions and four each in two divisions. The most useful breakdown seems to be that by Early, Middle, and Late Holocene - in this time series there were 37, 16, and 36 papers respectively, indicating best coverage to be of the Late and Early Holocene, and least well covered to be the Middle Holocene (6600-3400 yr B.P.).

Upon completion of this project, further analysis of the information depicted in Tables 6 and 7, and Figure 1 could prove quite useful in research planning.

SUMMARY

The Annotated Bibliography of Holocene Paleoclimates Project is described as to methods of: (1) selection of materials for inclusion; (2) formatting of the bibliographic records, including arrangement into categories based on method of climatic reconstruction, information fields to be displayed in the published product, indexes, maps, and a time table; and (3) computerization of production of the bibliography through input into the TRS 80 III, transfer to the IBM 3033 at the Carbon Dioxide Information Center at Oak Ridge National Laboratory, and computer publication and printing of the final work, complete with computer-produced indexes. Over one-third of the estimated final total of some 1200 records are currently available, and are used to illustrate the types of information that can be obtained through analysis of the groupings produced under various types of indexing terms. This preliminary example is used to demonstrate coverage of the literature at various access points (method of climatic reconstruction, area, and time), and to show a little bit about the adequacy of coverage in the literature of the "critical periods" in time important to the research for the Climatic Change in Canada Project. Upon initial observation, it is evident that the period around 6000 yr B.P. lacks the amount of coverage present for the other time periods.

ACKNOWLEDGEMENTS

This project is being supported by the United States Department of Energy, Carbon Dioxide Research Division, through contract DE-AC02-81ER60011 with the Regents of the University of Colorado. The bibliographic file resulting from this project will be merged with the United States Department of Energy Carbon Dioxide Information Center Bibliographic Information System

TABLE 7: *INDEX TO GEOGRAPHIC DESCRIPTORS.*

(Number beside name indicates number of times descriptor has been used; each paper is indexed to only one area, 465 papers have been indexed.)

Alaska	2
Alaska, Central	5
Alaska, North Slope	13
Alaska, Southeast	11
Alberta	4
Aleutian Islands and surrounding waters	8
Alps	8
Andes	2
Antarctica	4
Antarctica, East	11
Antarctica, West	13
Arctic	14
Arctic Ocean	2
Argentina and Chile	3
Baffin Bay--Davis Strait	3
Baffin Island	14
Baffin Island, East	17
Baffin Island, South	2
Baffin Island, West	11
Bering Sea	3
British Columbia	7
Canada	13
Canadian Arctic Islands Waters	1
Carpathians	1
Cascade Range	2
Caucasus	1
Cevennes	2
Cordillera Oriental	1
Denmark	3
East Africa	2
Ellesmere Island	12
Europe	2
Finland	9
Franklin District	6
General (non-geographic)	7
Global	15
Greenland	4
Greenland, East	8
Greenland, Inland Ice	8
Greenland, North	2
Greenland, West	12
Himalayas	1
Hindu Kush	1
Hudson Bay--James Bay	4
Iceland	3
Kara Sea	1
Keewatin District	5
Labrador	9
Mackenzie District	2
New Zealand	6
North America	7
North Atlantic Ocean	7
North Pacific Ocean	1
Northern Hemisphere	12
Northwest Territories	4
Norway	18
Ontario, Northern	1
Polar Regions	6
Quebec	3
Queen Elizabeth Islands	13
Rocky Mountains	23
Scandinavia	6
Sierra Nevada	3
Snowy Mountains	1
Southern Hemisphere	1
Svalbard and Svalbard waters	10
Sweden	6
The Faeroes	1
Ungava	5
United Kingdom and Ireland	9
United States, Western	8
USSR	6
USSR, East	2
USSR, West	4
Yukon Territory	6

(CDIC/BIS) file currently online at Oak Ridge National Laboratory. I thank Dr. Charles Weisbin, Director of CDIC, and members of his staff for assistance in entering the "Annotated Bibliography of Holocene Paleoclimates" file.

I also thank the INQUA Paleoclimate Commission for backing the pilot study which preceded the successful proposal to DOE, for further financial aid, and especially for supporting the idea of the project.

Special thanks go to Dr. Alan Hecht, then of the Climatic Dynamics Division of the United States National Science Foundation, for his interest in the project and for his important suggestions concerning funding.

REFERENCES

Andrews, Martha. 1980. Report on pilot study of bibliographic data systems' usability for Holocene paleoclimate research. Unpublished report to INQUA Paleoclimate Commission. 7 pp. plus appendices.

Andrews, Martha, and J.T. Andrews. 1980. A proposal to the Department of Energy for support of an annotated bibliography of Holocene paleoclimate literature. (Unpublished). 25 pp.

APPENDIX 1: BIBLIOGRAPHIES USED AS SOURCES FOR MATERIALS FOR THE ANNOTATED BIBLIOGRAPHY OF HOLOCENE PALEOCLIMATES

Printed Indexes

Andrews, Martha, and J.T. Andrews. 1980. Baffin Island Quaternary environments: an annotated bibliography. Institute of Arctic and Alpine Research, University of Colorado, Occasional Paper No. 33:1-123.

Antarctic Bibliography. Prepared by Library of Congress: sponsored by Office of Antarctic Programs, National Science Foundation (volumes 1-3) and Office of Polar Programs, National Science Foundation (volumes 4-11). Edited by: George A. Doumani (volumes 1-2) and Geza Thuronyi (volumes 3-11). Washington, D.C., Library of Congress, 11 volumes (volume 1, 1965-volume 11, 1981).

Antarctic Bibliography. 1951-1961. Prepared by the Library of Congress and sponsored by the Office of Polar Programs, National Science Foundation. Washington, Library of Congress, 1970. 349 pp.

Arctic Bibliography. Prepared by the Arctic Institute of North America. Edited by: Marie Tremaine (volumes 1-14) and Maret Martna (volumes 15-16). Washington, D.C., Department of Defense (volume 1, 1953-volume 11, 1963); Superintendent of Documents (volume 12, 1965); Montréal and London, McGill - Queen's University Press (volume 14, 1968-volume 16, 1975).

Chilton, B.D., L.J. Allison, and S.S. Talmage. 1981. Global aspects of carbon dioxide: an annotated bibliography. Oak Ridge, Tennessee, Oak Ridge National Laboratory, ORNL/EIS-195. 229 pp.

Clayton, Keith M. 1966. Geomorphological Abstracts Index 1960-1965 (Nos. 1-27). London, Geo Abstracts, London School of Economics. 371 pp.

Clayton, Keith M., and Margaret A. Bass. 1972. Geographical Abstracts A - Geomorphology Cumulative Index 1966-1970. Norwich, England, Geo Abstracts. 1023 pp.

Clayton, Keith. 1979. Geo Abstracts A - Landforms and the Quaternary Cumulative Index 1971-1975. Norwich, England, Geo Abstracts. 1027 pp.

Ghanimé, Linda. 1979. An annotated bibliography of Quaternary climatic changes in Canada. (Preliminary draft). 287 pp.

Newson, M., J.P. Barkham, J. Ash, and K. Beven. 1980. Geo Abstracts B - Climatology, Hydrology and Biogeography Cumulative Index 1971-1975. Norwich, England, Geo Abstracts, 1051 pp. Biogeography covered 1971-1973 only.

Peterson, N. Merle. 1974-1981. Ecology of the Canadian Arctic Archipelago: selected references. Department of Indian Affairs and Northern Development, Ottawa. 9 volumes.

Yates, E.M., and Margaret A. Bass. 1972. Geographical Abstracts B - Biogeography and Climatology, Cumulative Index 1966-1970. Norwich, England, Geo Abstracts. 954 pp.

Computerized Databases[1]

BIOSIS PREVIEWS, 1969-present, 3,169,540 records, semi-monthly updates (Biosciences Information Service, Philadelphia, PA).

---

[1] Commercially available through DIALOG Information Services, Inc., if availability is not listed at end of citation.

APPENDIX 1: (Cont'd)

GEOARCHIVE, 1969-present, 336,000 citations, monthly updates (Geosystems, London, England).

GEOREF, 1961-present (North American material), 1967-present (worldwide material), 677,500 records, monthly updates (American Geological Institute, Falls Church, VA 22041).

NTIS, 1964-PRESENT, 863,500 citations, biweekly updates (National Technical Information Service, U.S. Department of Commerce, Springfield, VA).

OCEANIC ABSTRACTS, 1964-present, 124,200 records, bimonthly updates (Cambridge Scientific Abstracts, Bethesda, MD).

Carbon Dioxide Information Center Bibliographic Information System, 1980-present, over 2100 records (Carbon Dioxide Information Center, Oak Ridge National Laboratory, Oak Ridge, TN 37830). Availability information upon request.

SPIRES/Boreal, catalogued collection of the Boreal Institute of Northern Studies Library (University of Alberta, Edmonton, Alberta). In-house system.

SPIRES/ASTIS, Arctic Science and Technology Information System, 1978-present, over 7000 records (Arctic Institute of North America, University of Calgary, Calgary, Alberta). Available by subscription.

QL/ASFA, Aquatic Sciences and Fisheries Abstracts, 1978-present, 98,179 documents (QL Information Systems, Canada). Some availability in United States?

ANNOTATED BIBLIOGRAPHY OF QUATERNARY CLIMATIC CHANGE IN CANADA, AND A BRIEF ANALYSIS OF ITS CONTENTS

C.R. Harington[1]

INTRODUCTION

Since its beginning in 1977, a goal of the National Museum of Natural Sciences Climatic Change Project was to compile a comprehensive, annotated bibliography on climatic change in Canada during the Quaternary (Harington 1980, p. 13). The work has been completed and will be published in 1984 as "Climatic Change in Canada 4" (*Syllogeus* 51). The work has 368 pages and contains 912 annotations. The latest references in the volume date to early 1983. It is hoped that the bibliography will: (1) serve as a "refresher" on sources, as well as being a useful research tool for professionals; (2) act as an introduction to references on climatic change in Canada during the last 2 million years (Quaternary) for laymen and students interested in the subject and for scientists beginning work in this field; (3) give some idea of the diversity of disciplines contributing published paleoclimatic data; and (4) demonstrate the acceleration of research in this field during the last few decades, as well as indicating some strengths and weaknesses of such research, where geographical, geochronological and subject coverage is concerned.

Fundamental to this work is a belief that a knowledge of our past climate, if carefully accumulated and interpreted, can lead to a better understanding of the nature, timing and strength of future climatic alterations. Regardless of predictive value, I think that the results of past work on Quaternary climatic change in Canada, only imperfectly indicated in this bibliography, will be of interest and value academically (e.g. to biogeographers, paleoecologists, archaeologists and historians).

PLANNING AND EXECUTION

Most of the work on this bibliography was carried out by two graduate students, Linda Ghanimé and Anne Smithers, working on contract under my supervision. In 1979, Linda Ghanimé compiled about 500 annotations - mainly from libraries in Montréal. Prior to her work, I did not realize how diverse and abundant references were on this subject, nor how far back in time of publication they extended.

---

[1] Paleobiology Division, National Museum of Natural Sciences, National Museums of Canada, Ottawa, Ontario K1A 0M8

Anne Smithers worked on the bibliography during two periods from 1981 to 1983. She greatly augmented and improved the bibliography by: (1) cross-checking reference lists in the most important papers to see if sources were being missed; (2) making annotations of pertinent references found in the GEOREF computer listing for 1961-1981 that applied to this topic[1]; (3) updating the annotations by checking current issues of the main journals containing papers on Quaternary paleoclimates in Canada (e.g. *Arctic*, *Arctic and Alpine Research*, *Canadian Journal of Earth Sciences; Climatic Change* and *Palaeogeography*, *Palaeoclimatology*, *Palaeoecology*); (4) renumbering the annotations that were finally selected and listing key words for each; and (5) preparing the index. Anne's main sources of information were libraries in Ottawa.

I was responsible for planning the bibliography, supervising the work of the contractors, selecting annotations for inclusion, contributing some annotations, organizing the index (with valuable advice from Martha Andrews, Cynthia Wilson and Anne Smithers), and, with help from Gail Rice, for editing the text.

ORGANIZATION

Annotations

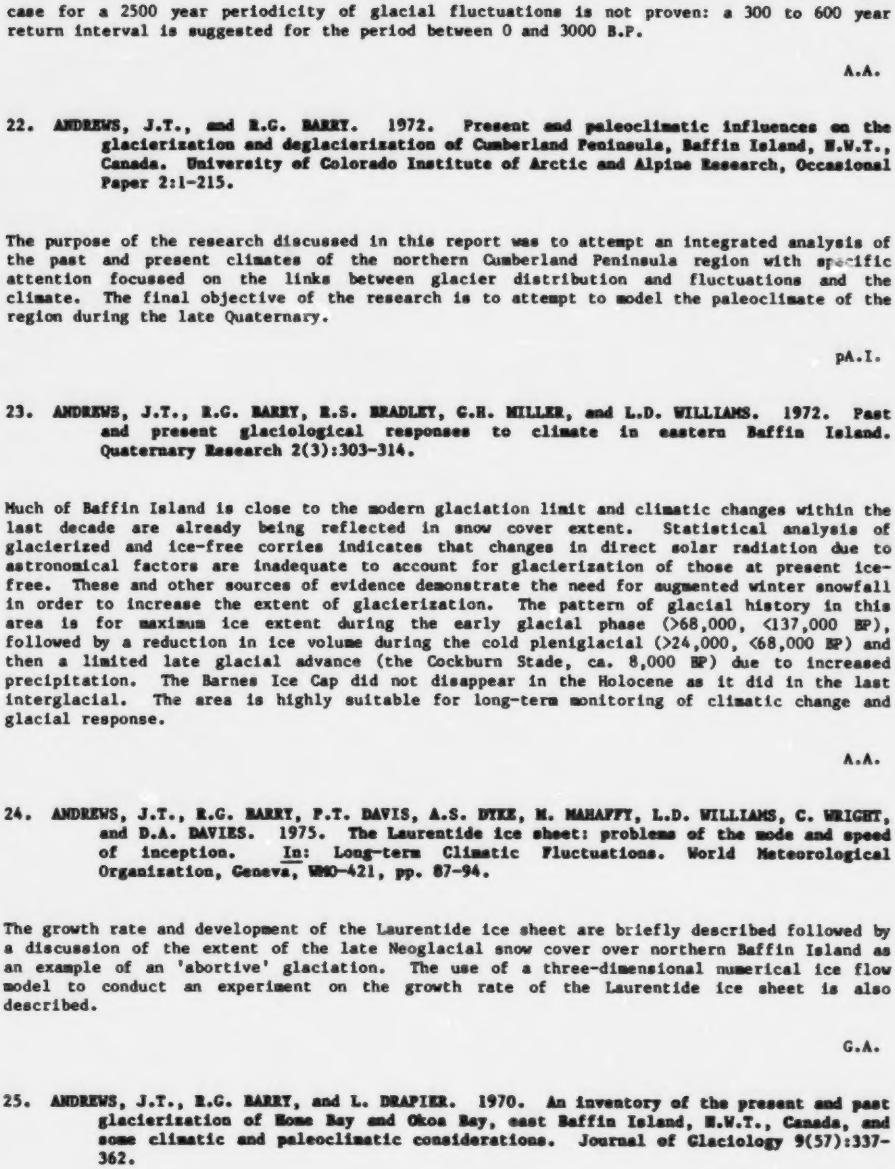
Annotations are listed in alphabetical order by author's (or first author's) surname and date of publication. Each annotation is preceded by a number for ready reference and to make the index at the end of the volume more succinct. Abbreviations following the annotations indicate their sources (e.g. A.A. - Author's abstract; G.A. - Abstract taken from Geo Abstracts; L.G. - Abstract prepared by Linda Ghanimé). For example see Figure 1.

Index

The index is organized by: (1) **Geographic region** - alphabetically according to province and territory names ("Northwest Territories" is divided into "Mainland" and "Islands" categories for ease of reference, and the regions (I to XIII) are marked on a map; Figure 2),

---

[1] The most useful key words in searching were: Canada (or province or territory names); Quaternary; Pleistocene; Holocene; and paleoclimatology.



case for a 2500 year periodicity of glacial fluctuations is not proven: a 300 to 600 year return interval is suggested for the period between 0 and 3000 B.P.

A.A.

22. **ANDREWS, J.T., and R.G. BARRY. 1972. Present and paleoclimatic influences on the glacierization and deglacierization of Cumberland Peninsula, Baffin Island, N.W.T., Canada. University of Colorado Institute of Arctic and Alpine Research, Occasional Paper 2:1-215.**

The purpose of the research discussed in this report was to attempt an integrated analysis of the past and present climates of the northern Cumberland Peninsula region with specific attention focussed on the links between glacier distribution and fluctuations and the climate. The final objective of the research is to attempt to model the paleoclimate of the region during the late Quaternary.

pA.I.

23. **ANDREWS, J.T., R.G. BARRY, R.S. BRADLEY, G.H. MILLER, and L.D. WILLIAMS. 1972. Past and present glaciological responses to climate in eastern Baffin Island. Quaternary Research 2(3):303-314.**

Much of Baffin Island is close to the modern glaciation limit and climatic changes within the last decade are already being reflected in snow cover extent. Statistical analysis of glacierized and ice-free corries indicates that changes in direct solar radiation due to astronomical factors are inadequate to account for glacierization of those at present ice-free. These and other sources of evidence demonstrate the need for augmented winter snowfall in order to increase the extent of glacierization. The pattern of glacial history in this area is for maximum ice extent during the early glacial phase (>68,000, <137,000 BP), followed by a reduction in ice volume during the cold pleniglacial (>24,000, <68,000 BP) and then a limited late glacial advance (the Cockburn Stade, ca. 8,000 BP) due to increased precipitation. The Barnes Ice Cap did not disappear in the Holocene as it did in the last interglacial. The area is highly suitable for long-term monitoring of climatic change and glacial response.

A.A.

24. **ANDREWS, J.T., R.G. BARRY, P.T. DAVIS, A.S. DYKE, M. MAHAFFY, L.D. WILLIAMS, C. WRIGHT, and D.A. DAVIES. 1975. The Laurentide ice sheet: problems of the mode and speed of inception. In: Long-term Climatic Fluctuations. World Meteorological Organization, Geneva, WMO-421, pp. 87-94.**

The growth rate and development of the Laurentide ice sheet are briefly described followed by a discussion of the extent of the late Neoglacial snow cover over northern Baffin Island as an example of an 'abortive' glaciation. The use of a three-dimensional numerical ice flow model to conduct an experiment on the growth rate of the Laurentide ice sheet is also described.

G.A.

25. **ANDREWS, J.T., R.G. BARRY, and L. DRAPIER. 1970. An inventory of the present and past glacierization of Home Bay and Okoa Bay, east Baffin Island, N.W.T., Canada, and some climatic and paleoclimatic considerations. Journal of Glaciology 9(57):337-362.**

An air-photograph inventory of the present glacierization of areas of east Baffin Island adjoining Home Bay and Okoa Bay is described. Ice fields characterize the broad mountain

FIGURE 1: *Sample page (p. 22) from "Climatic Change in Canada 4 Annotated Bibliography of Quaternary Climatic Change in Canada"* (Syllogeus *51*).

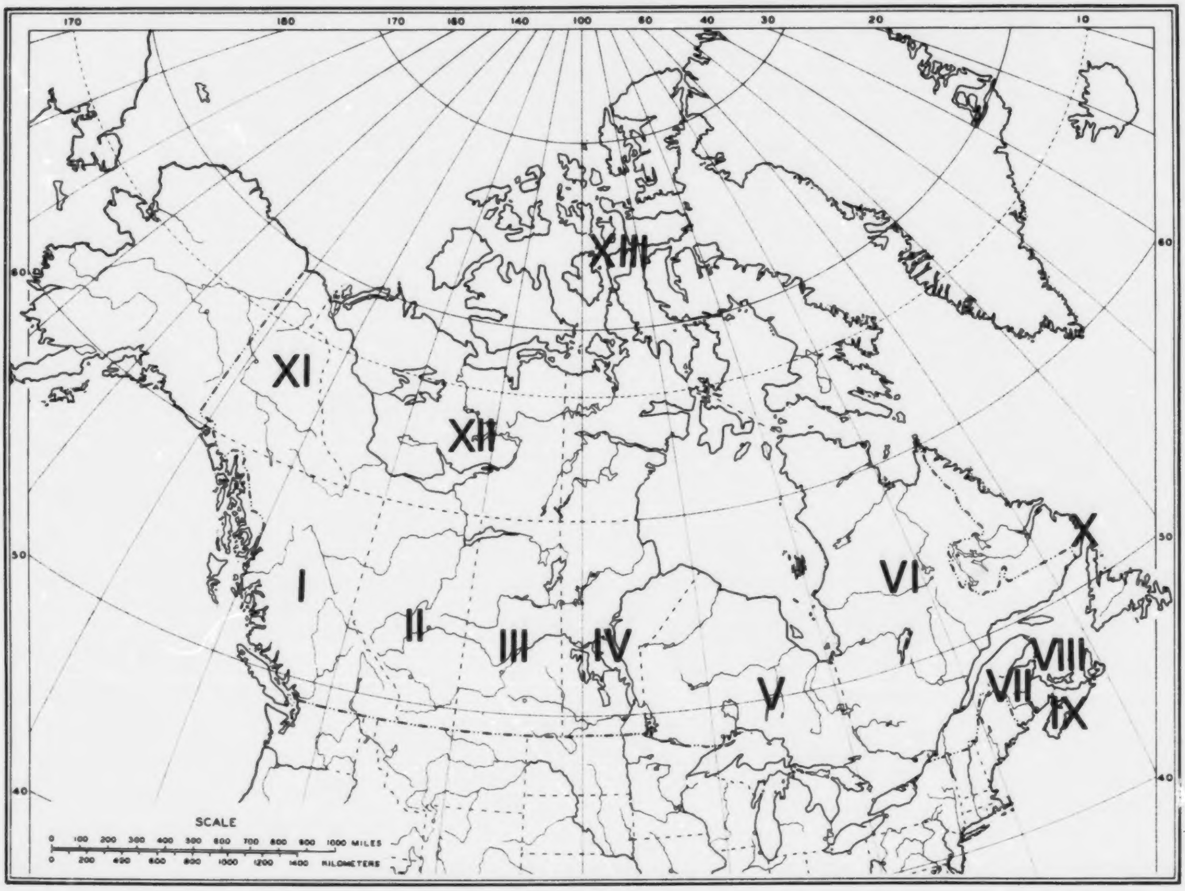

170
180
170
160
140
100
80
40
30
20
10
60
50
40
XIII
XI
XII
X
VI
I
II
III
IV
V
VIII
VII
IX
SCALE
0 100 200 300 400 500 600 700 800 900 1000 MILES
0 200 400 600 800 1000 1200 1400 KILOMETERS

FIGURE 2: *Geographical regions of Canada used in the index; I - British Columbia; II - Alberta; III - Saskatchewan; IV - Manitoba; V - Ontario; VI - Québec; VII - New Brunswick; VIII - Prince Edward Island; IX - Nova Scotia; X - Newfoundland; XI - Yukon Territory; XII - Northwest Territories (Mainland); XIII - Northwest Territories (Islands).*

"Canada-General" (where annotations deal with Canadian areas larger than natural regions), "North America" (annotations concerning Canada and other parts of North America), and "Global" (annotations dealing with Canada and other parts of the world); (2) **Geological time** - "Quaternary" (approximately the last 2 million years), "Pleistocene" (approximately 2 million years ago to approximately 10,000 yr B.P.), "Holocene" (approximately the last 10,000 years), "Holocene - Prehistoric" (approximately 10,000 yr B.P. to approximately 500 yr B.P.), and "Holocene - Historic" (approximately 500 yr B.P. to the present) - Figure 3; (3) **Subjects** - nearly 40 names of disciplines or topics are listed alphabetically following the "General" category. Some subjects (e.g. "Climate" and "Geology") are subdivided for greater precision of reference (Table 1).

ANALYSIS OF BIBLIOGRAPHIC DATA

In addition to publishing the bibliography for use as a research tool and for educational purposes, perhaps it is worthwile to briefly review its contents in an effort to answer some questions about the status of Quaternary paleoclimatic research in Canada.

1. What are the earliest papers that deal specifically with Quaternary paleoclimate in Canada?

   Answer: Dawson, Sir J.W. 1893. The Canadian Ice Age: being notes on the Pleistocene Geology of Canada, with special reference to the life of the period and its climatal conditions. W.V. Dawson, Montreal. 301 pp., and Coleman, A.P. 1895. Glacial and interglacial deposits near Toronto. Journal of Geology 3:622-645.

2. Has the number of published papers concerning Quaternary climate in Canada increased substantially over the past 10 or 20 years?

   Answer: Yes (see Figure 4). Papers begin to appear in the 1890s, and relatively few are published until the 1950s. There has been an acceleration of publishing in this field from 1946 to 1980, reaching 280 papers over the period 1976-1980.

3. What subjects have received most attention in the field of Quaternary paleoclimatic research in Canada?

   Answer: Palynology (13.3%), Geology (12.3%), Paleoecology (11.5%), and Glaciology (8.7%). In contrast, research based on Historic Accounts (1.0%) and Agriculture (0.3%) has received comparatively little attention.

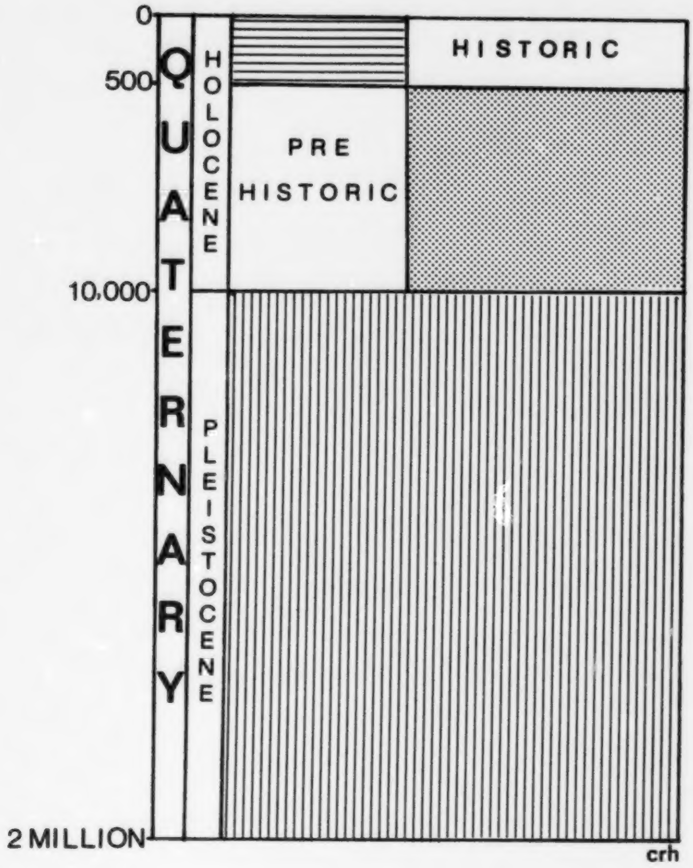

0
500
10,000
2 MILLION
QUATERNARY
HOLOCENE
PLEISTOCENE
PRE HISTORIC
HISTORIC
crh

FIGURE 3: *Geological time periods used in the index. Numbers on left represent years B.P. (approximate).*

4. What regions have received the best coverage?

   Answer: Northwest Territories (Islands) (16.7%), Québec (16.1%) and Ontario (11.8%) are best covered. Alberta (8.2%), British Columbia (8.1%), Northwest Territories (Mainland) (7.7%), Newfoundland (7.3%), Yukon Territory (5.5%) and Nova Scotia (5.4%) are relatively well covered. Manitoba (4.8%), Saskatchewan (3.8%), New Brunswick (2.7%) and Prince Edward Island (1.9%) have received relatively little attention.

5. What periods of geological time have received the best coverage?

   Answer: Holocene - Historic (approximately 500 yr B.P. to the present) (37.2%) and Holocene - Prehistoric (approximately 10,000 yr B.P. to approximately 500 yr B.P.) (34.1%).

In future, I suggest it would be useful for professionals in various subjects to analyze in detail and synthesize information available on Quaternary paleoclimate in the different geographic regions of Canada. An example of this approach is Chapter 6, "Paleoenvironmental Reconstruction" in Ritchie's (1984) book "Past and Present Vegetation of the Far Northwest of Canada". Such syntheses could also be carried out on a nationwide basis for the various critical periods dealt with in this volume.

SUMMARY

"The Annotated Bibliography of Quaternary Climatic Change in Canada" will be published in 1984. It contains 912 annotations, which are listed alphabetically by author's surname and date of publication, and are indexed according to geographical region, geological time and subject (nearly 40 subjects are used). This is a preliminary attempt at establishing a source of substantial information on mainly published papers dealing with climatic change and variability in Canada during the last 2 million years.

A brief analysis of the contents of the bibliography indicates that: (1) the earliest paper specifically mentioning Quaternary paleoclimate in Canada is Sir J.W. Dawson's (1893); (2) there has been a rapid increase in the publication of papers on the subject since 1946-1950 (according to Figure 4, publication declined during the Second World War - possibly due to deflections of manpower and economic support for military purposes); (3) palynology, geology, paleoecology and glaciology have received the best coverage of all subjects listed in the bibliography; (4) as far as geographic areas are concerned, Northwest Territories

TABLE 1: *SUBJECTS LISTED IN THE INDEX*

General

Agriculture

Archaeology (including Culture)

Atmospheric (including Circulation, Influences)

Bibliographies

Chemistry (including Isotopic Analysis)

Climate - General
    - Trends
    - Forecasting
    - Modelling

Concepts

Dating Methods (including Radiocarbon Dating)

Fire History

Foraminifera

Geology - General
    - Geomorphology
    - Stratigraphy
    - Sedimentology

Glaciology

Historic Accounts

Hydrology

Insects

Lichenometry

Molluscs

Ostracodes

Oceanography

Paleobotany (including Plant Macrofossils)

Paleoecology

Paleomagnetism

Paleoceanography

Palynology

Pedology (including Paleosols)

Sea Level

Solar Effects

Speleology

Temperature Reconstructions

Tree-ring Studies (including Dendrochronology and Dendroclimatology)

Vertebrates

Volcanic Ash (including Tephrochronology and Tephrostratigraphy)

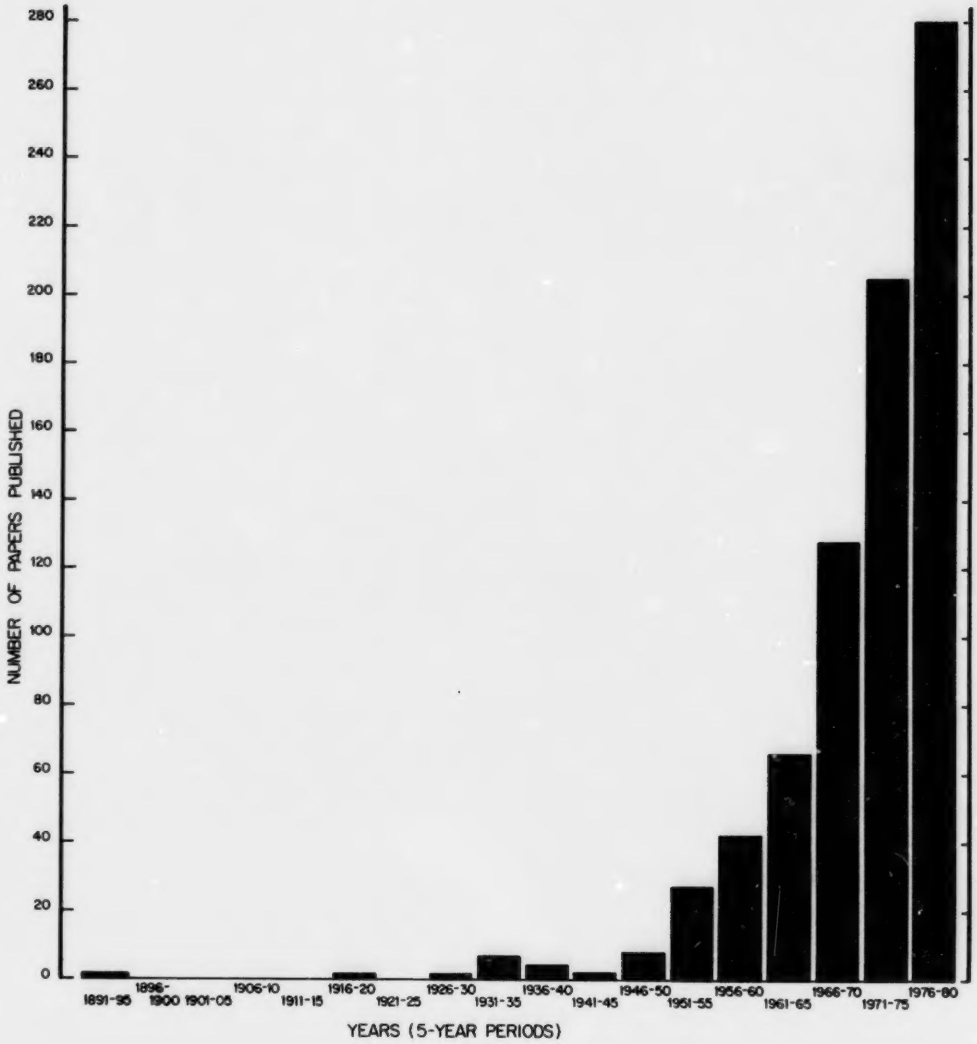


FIGURE 4: *Numbers of papers published over 5-year periods concerning Quaternary climate in Canada (1891-1980).*

(Islands), Québec and Ontario have received the most thorough treatment; (5) the Holocene has received much better coverage than the Pleistocene, which is natural seeing that the latest materials tend to be best preserved and most readily accessible for study.

A remarkable fact emerging from the foregoing analysis is the relative abundance of references mentioning Quaternary paleoclimate in the Canadian Arctic Islands. This was first brought to my attention by Anne Smithers, and has also been noted by Martha Andrews (this volume). The main reason appears to be that much of this work stemmed from long-range, interdisciplinary studies on Baffin Island by research teams from the Institute of Arctic and Alpine Research at Boulder, Colorado, from similar projects of the Terrain Sciences Division, Geological Survey of Canada, and of the Polar Continental Shelf Project.

At this stage, it would be useful to have careful syntheses of Quaternary paleoclimatic evidence for various regions, as well as nationwide syntheses of such evidence for the critical periods mentioned in this volume.

ACKNOWLEDGEMENT

I thank Jennifer Dawson for help with statistical analysis of the contents of the bibliography.

REFERENCES

Harington, C.R. 1980. The impact of changing climates on people; and the National Museum of Natural Sciences climatic change project. In: Climatic Change in Canada. Edited by: C.R. Harington. Syllogeus No. 26:5-15.

Ritchie, J.C. 1984. Past and present vegetation of the Far Northwest of Canada. University of Toronto Press, Toronto. 251 pp.

THIRTIES DROUGHT ON THE PRAIRIES - HOW UNIQUE WAS IT?

M.O. Berry[1] and G.D.V. Williams[1]

INTRODUCTION

During the past century the climatic event which perhaps has been most critical to Canadian society was the 1930s prairie drought. It resulted in direct losses to wheat production estimated at about two billion (1980) dollars. Losses to other types of farming and other parts of the economy reliant on agriculture must have increased the total greatly. In more human terms, it left a substantial portion of the prairie population with the choice of either emigrating or living in abject poverty. Of course the troubles of this period were not restricted to drought. Damage to crops from pests and diseases played a role. The Depression had a major effect on the economy. However, for the prairie farmer it was probably the lack of water and resulting crop losses which more than anything else disrupted his life. The Relief Commission in Saskatchewan, the hardest hit province, commented at the time that "while the other provinces were suffering from a world-wide depression, Saskatchewan was, in addition, experiencing consecutive crop failures over an extensive area as a result of one of the worst droughts in history" (Britnell 1939).

In this paper, we examine the nature of the climatic events of this period, and make a preliminary assessment of their uniqueness - in other words the probability of their recurrence. An objective of the work on which much of this paper is based is to provide information needed by individuals and agencies concerned with planning for the future. We also attempt to demonstrate the limitations placed on this type of analysis by the use of data from only the period in which a fairly dense regular weather observing network exists.

---

[1] Atmospheric Environment Service, Downsview, Ontario M3H 5T4

DATA

Extremely important in the assessment of the probability of recurrence of a particular climatic event, such as drought, is the availability of a sufficiently long period of reliable weather data. It is, therefore, useful to consider some of the information available from the prairie region, as follows:

1920s-1980s	Numerous network stations.
1870s-1920s	Some stations with regular observations.
1800s-1870s	Fragmented instrumental observations, diaries, etc.
1600s-1900s	Tree rings.

No doubt other types of information would also help reconstruct past climate back for a number of centuries. A question mark should follow tree-ring data because it is, at present, not clear that this source will yield much useful information about prairie droughts.

As mentioned above, for a number of reasons it is only the first category of data, from 1928 to 1980, which is used our subsequent analysis.

METHOD

A good starting point for a drought study is the establishment of exactly what one is going to study. What do we mean by the term drought? Many definitions have been developed, however for purposes of this paper, it will be defined as a shortage of soil moisture sufficient to cause a major decrease in wheat yields, with "major" being defined rather arbitrarily, as will be seen.

The grid illustrated in Figure 1, with points separated by 100 km, was used for the study. To it were transferred temperature and precipitation data from all suitable observing stations, for a period from the 1920s to 1980. Soil and crop information were also assigned to grid points. By incorporating water budget methodology developed by Palmer (1965), Baier *et al.* (1979) and the United States Environmental Protection Agency (1980), a model was developed to simulate the hydrological cycle, as illustrated in Figure 2. This model was used with the grid-point data to calculate soil moisture at two levels, the first extending down to approximately 15 cm below the surface, and a second, deeper layer. Time series consisting of values calculated at 10-day intervals were prepared, extending back to

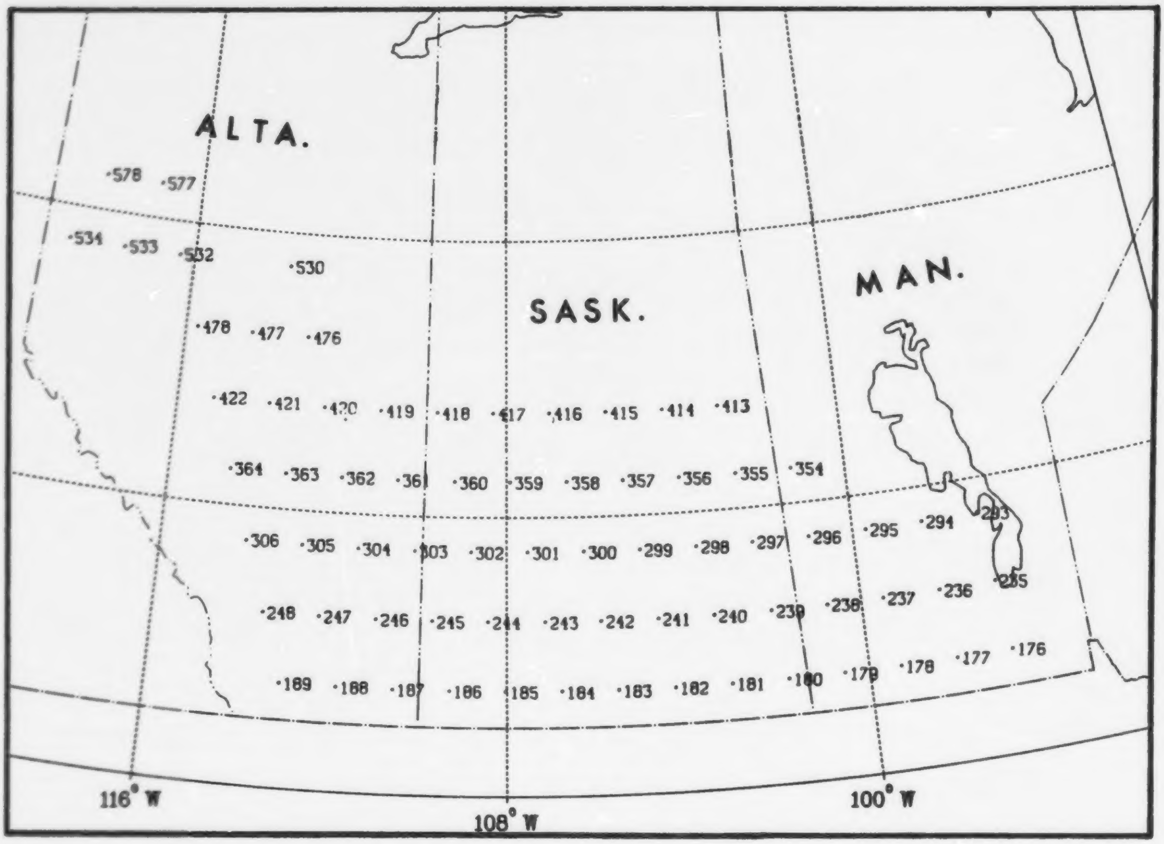

ALTA.
SASK.
MAN.
·578 ·577
·534 ·533 ·532 ·530
·478 ·477 ·476
·422 ·421 ·420 ·419 ·418 ·417 ·416 ·415 ·414 ·413
·364 ·363 ·362 ·361 ·360 ·359 ·358 ·357 ·356 ·355 ·354
·306 ·305 ·304 ·303 ·302 ·301 ·300 ·299 ·298 ·297 ·296 ·295 ·294 ·293
·248 ·247 ·246 ·245 ·244 ·243 ·242 ·241 ·240 ·239 ·238 ·237 ·236 ·235
·189 ·188 ·187 ·186 ·185 ·184 ·183 ·182 ·181 ·180 ·179 ·178 ·177 ·176
116° W
108° W
100° W

FIGURE 1: *Grid points used in data analysis.*

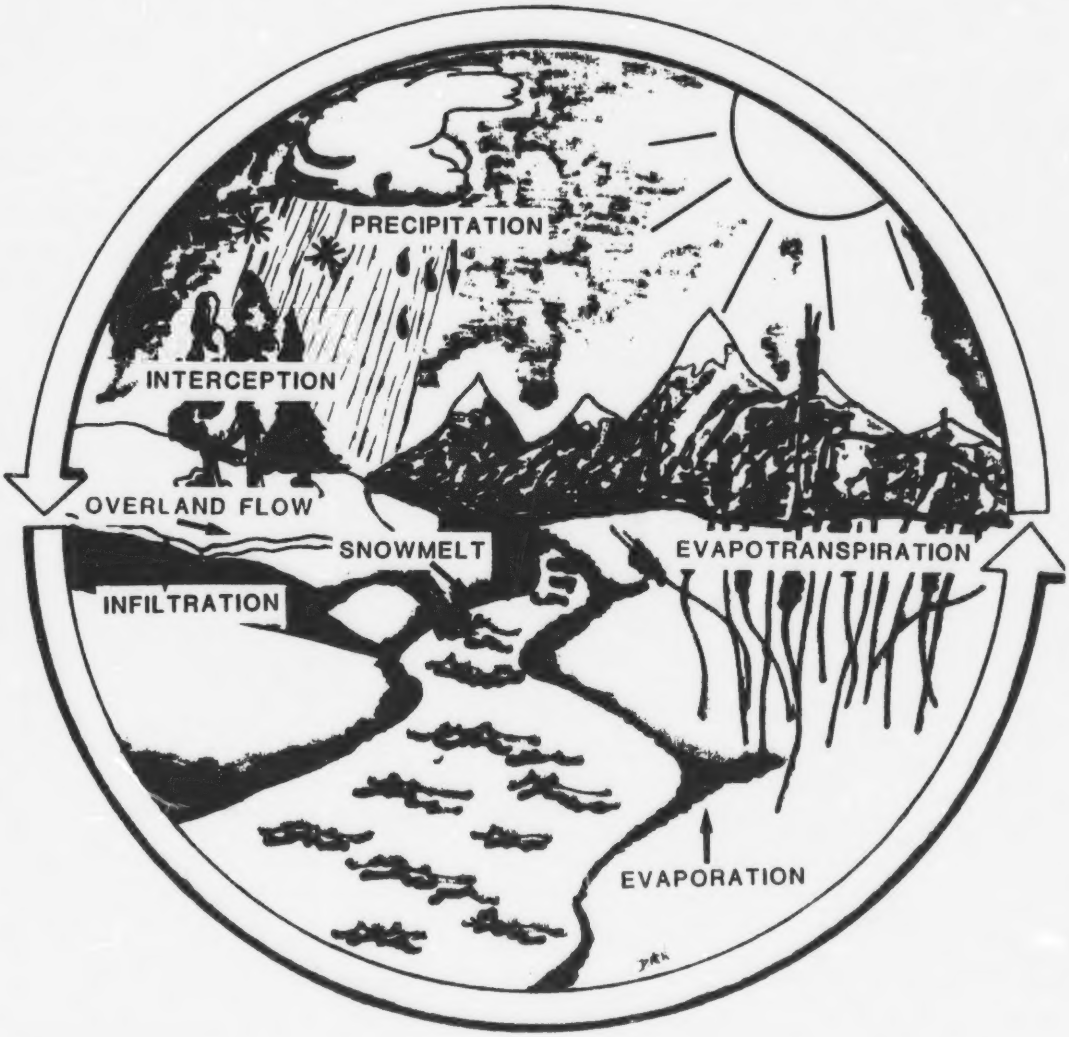

PRECIPITATION
INTERCEPTION
OVERLAND FLOW
SNOWMELT
EVAPOTRANSPIRATION
INFILTRATION
EVAPORATION
DRK

FIGURE 2: *Processes of the hydrological cycle simulated by the water-budget program.*

the 1920s at most grid points. Empirical relationships were then developed between the soil-moisture values and yearly wheat yields. These relations were used to determine the year-to-year variation in yield associated with variations in the available soil moisture, which in turn is determined by weather conditions - particularly precipitation. The objective of developing these relationships was to distinguish changes in yield caused by moisture deficit from those associated with other factors such as improved agricultureal techniques, pests and diseases. For a more detailed discussion of the methodology, the reader should consult Street and Findlay (1981) and Williams (1983).

RESULTS

The estimates of variation in wheat yield associated with variations in available water have been termed Water-Based Yield Estimates (WBYE). Figure 3 displays values calculated on a kg/ha basis for each year of the study period, assuming a 1979 level of agricultural technology through the period. The average yield is 1820 kg/ha. A value of 1680 has been chosen, somewhat arbitrarily, as the threshold below which a major drought occurs on the Canadian Prairies. In the 53-year period shown in the figure, occurrences of these droughts are confined to a 10-year interval beginning in 1929 and a 12-year interval starting in 1957. The most severe year in the period does not occur in the 1930s, but rather in 1961. However, 1936 and 1937 rate second and third in severity.

It is interesting to note that in addition to (or perhaps as a result of) being grouped in the 1929-38 and 1957-68 periods, the major drought episodes apparently have a tendency to occur in pairs, with six of the nine identified falling into this category. Since the primary reason for doing this analysis was to determine the frequency and nature of future droughts, the question of the extent to which these features are real and persistent characteristics of prairie climate is important, particularly in the case of drought where several occurrences over a short period have considerably greater impact than the same number occurring at more widely-spaced intervals.

Given our present limited ability to model atmospheric processes, the answer to this question must be obtained through empirical means, i.e. through the statistical analysis of time series such as that in Figure 4, which depicts the probability of yields exceeding various amounts. Any point on the diagonal line in this figure would join horizontal and vertical lines connecting a yield whose probability of being exceeded is given by the

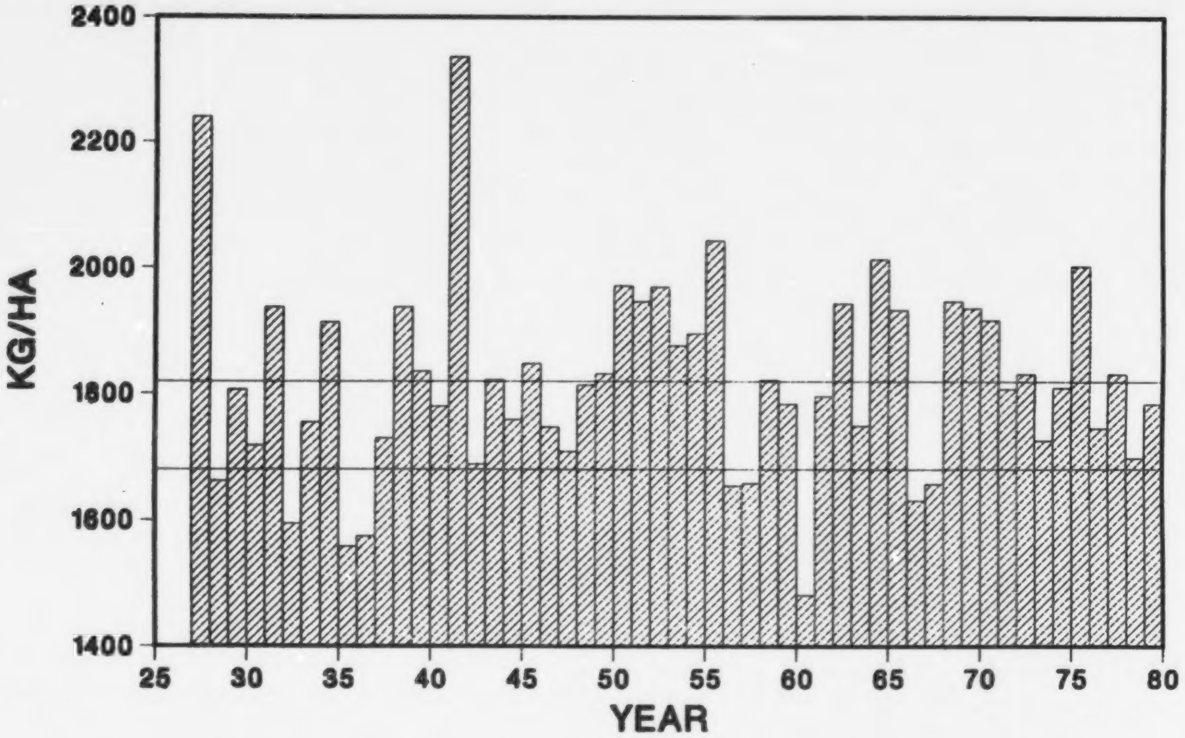


FIGURE 3: *Water-based yield estimates for western Canada (1927-1980). Values below lower line (1680 kg/ha) are classed as major droughts.*

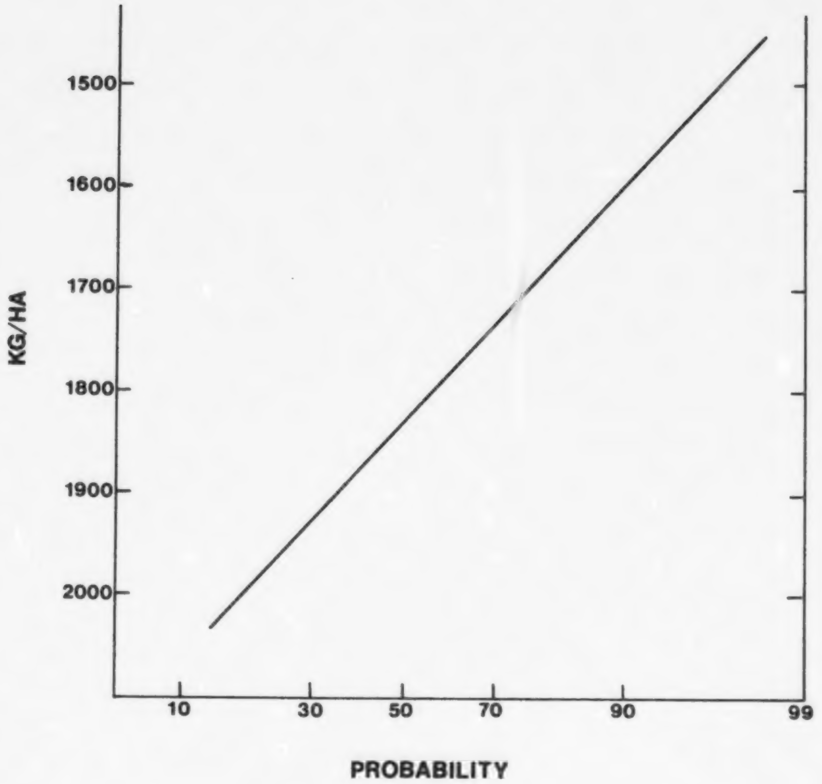


FIGURE 4: *Probability of prairie water-based yield estimates exceeding given values.*

corresponding value on the x-axis. It indicates that conditions as severe as those that occurred in 1936 - the year identified as most severe in the Thirties - have a probability of about 5% of occurring in any given year. Conditions sufficiently dry to have a major impact on production of wheat and other crops, as defined by the 1680 kg/ha threshold, have an occurrence probability of about 20%, or one year in five if separated by equal intervals. In other words, the analysis indicates that seriously dry conditions are not that infrequent, even to the point where lack of water is as serious as in the mid-Thirties.

This recurrence analysis is preliminary in nature, and needs to be refined and expanded. For example, it deals with the Canadian Prairies as a whole, but does not represent the risk to an individual farmer or locality, since the probability varies from point to point through the region, and since even the most severe droughts do not normally encompass the whole region. The latter situation is illustrated in Figure 5, which depicts estimated soil moisture in one of the driest periods (July 1937) in the 53 years studied. Although dry conditions are extensive, much of southern Manitoba and parts of central Alberta had near normal conditions.

It should also be noted that the analysis gives average recurrence intervals but does not provide any information on the probability that dry years will fall in groups, as suggested by the two series of dry years in Figure 3. Unfortunately, when dealing with events such as these, which appear to cluster at intervals 20 to 30 years apart, the 53 years of instrumental observations upon which this time series are based are insufficient to determine the extent to which these patterns may be expected to recur. Proxy data could play an important role in extending the series of climatological records back in time over periods when instrumental observations were scarce or nonexistent.

Figure 6 sums by decade the yield losses given in Figure 3. The losses estimated for the 1930s are the greatest of the five decades given, however they only slightly exceed those of the 1960s. Losses in the other three decades were much less. The Sixties do not bring to mind the hardship and disruption of lives that occurred in the Thirties. While the impacts of drought in the latter period have been extensively documented, analysis of the Sixties is rather limited. However, it seems likely that factors such as the higher degree of integration of the economy, availability of a much better economic safety net in the form of various government programs, and the lack of a concurrent economic depression must have been factors lessening individual hardship.

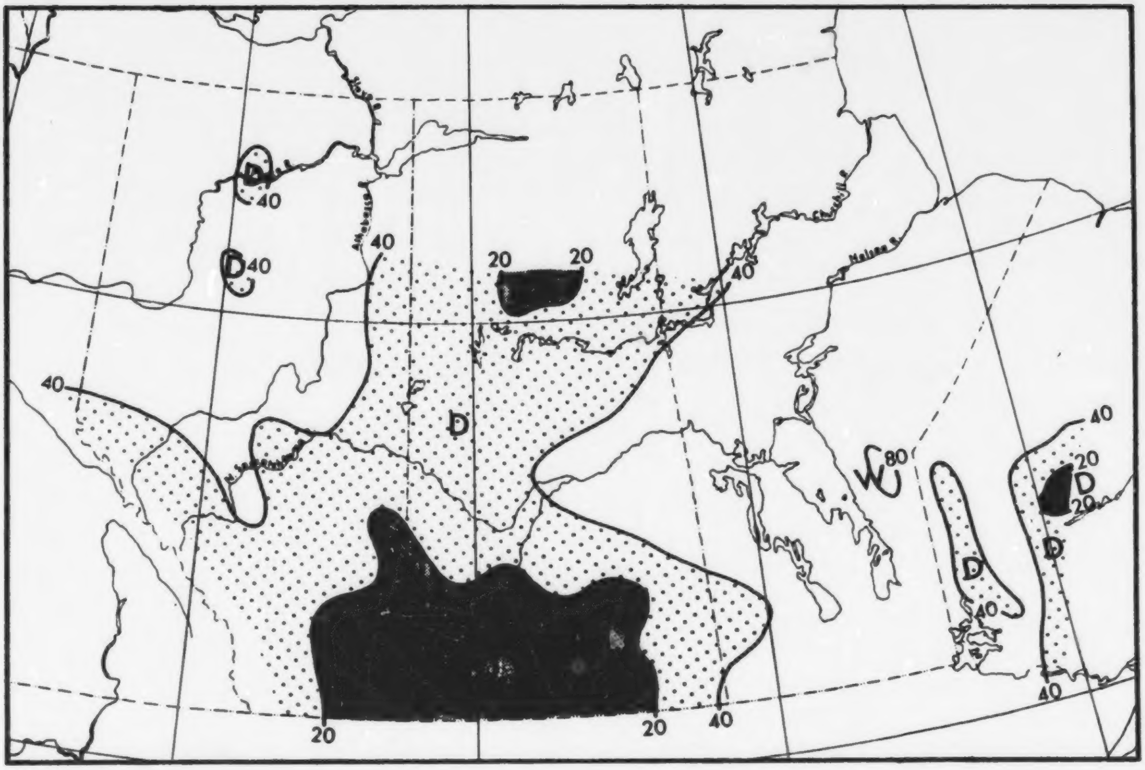

D
40
D
40
Athabasca R.
40
20
20
40
Churchill R.
Nelson R.
40
D
N. Saskatchewan R.
W
80
40
20
D
20
D
D
40
20
20
40
40

FIGURE 5: *Soil moisture, July 1937. "D" and "W" indicate dry and wet areas. Very dry areas are darkly shaded.*

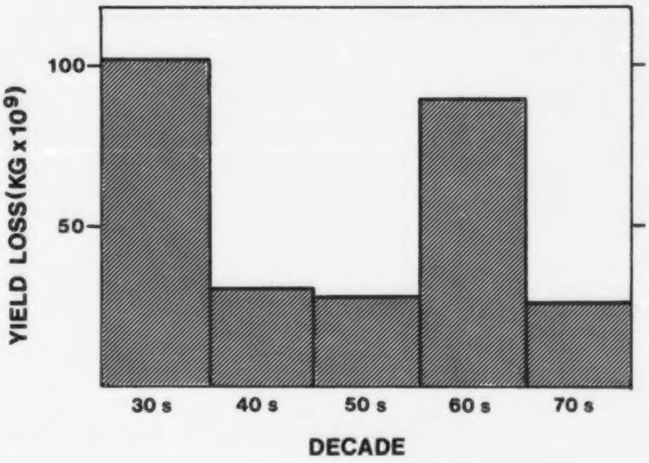


FIGURE 6: *Drought-related losses, prairie wheat.*

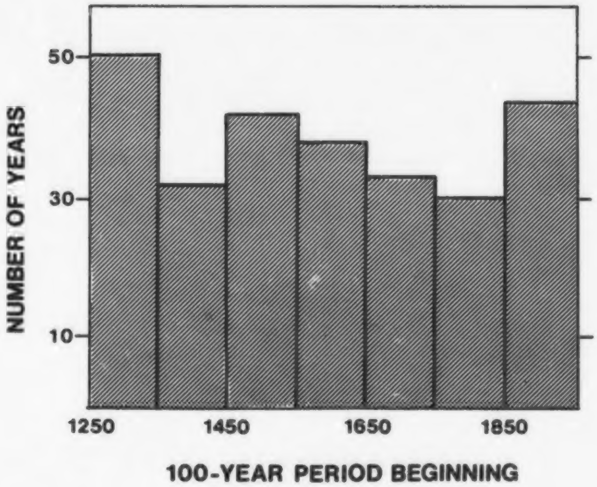


FIGURE 7: *Number of dry years per hundred-year period derived from tree-ring data (Weakly 1965).*

LONG-TERM STABILITY

Analyses of this type, which use climatic data from the past to predict frequencies and other characteristics of future events, normally are based on the assumption that climate will remain stable. Here again proxy data may aid the analyst by providing a long series which can be used to assess stability up to the present. Of course this does not ensure that conditions in the future will continue the same, but the data can provide useful evidence. For example, Figure 7 provides an index of dryness derived from tree rings and spanning a number of centuries. A subjective assessment of the 100-year average values presented in the diagram suggests a lack of any trend in precipitation over the period, but a considerable variation from one interval to the next. A more detailed analysis of the data would be needed before anything further could be said. In view of the fact it is derived from trees located at a considerable distance from the Canadian Prairies, the applicability of the information to Canada is doubtful.

Proxy data may also provide analogues for predicted changes in climate caused by anthropogenic or other effects. For example, a warm spell from the past could be used as a model for a warmer climate associated with increased levels of atmospheric carbon dioxide.

SUMMARY

Because this is a preliminary analysis and limited by the length of the time series of weather observations used, it cannot provide a complete answer to the question of how unique the 1930s prairie drought was. Yet it indicates that drought is not that uncommon an event, with conditions as dry as the worst year of the 1930s to be expected with an average frequency of once in 20 years. Droughts of lesser severity, but still substantial impact on grain production, would be expected to occur at greater frequencies.

If proxy data can be used to lengthen substantially the period for which climatic information is available, it would be of considerable benefit in climatological application. The need is demonstrated in the type of analysis described here, where the probability of future events, particularly extreme events, must be deduced from historic data. The greatest hurdle to achieving this appears not to be the obtaining of proxy data, but the development of an understanding of how climate affects them.

REFERENCES

Baier, W., J.A. Dyer, and W.R. Sharp. 1979. The versatile soil moisture budget. Agriculture Canada, Ottawa, Agricultural Meteorology Technical Bulletin 87:1-52.

Britnell, G.E. 1939. The wheat economy. University of Toronto Press, Toronto. 260 pp.

Palmer, W.C. 1965. Meteorological drought. United States Weather Bureau Research Paper No. 45:1-58.

Street, R.B., and B.F. Findlay. 1981. An objective climatological study of prolonged dry spells (meteorological drought) in the Prairie Provinces. Atmospheric Environment Service, Downsview. 29 pp.

United States Environmental Protection Agency. 1980. Hydrological simulation program. Environmental Research Laboratory, Athens, Georgia, EPA-600/9-80-015:1-678.

Weakly, H.E. 1965. Recurrence of drought in the Great Plains during the last seven hundred years. Agricultural Engineering 46:85.

Williams, G.D.V. 1983. Agricultural drought. (Chapter 4 of unpublished report on drought), Atmospheric Environment Service, Downsview.

THE EFFECTS OF MAJOR VOLCANIC ERUPTIONS ON CANADIAN CLIMATE

Walter R. Skinner[1]

INTRODUCTION

Two recent major volcanic eruptions, Mt. St. Helens in the United States in 1980 and El Chichon in Mexico in 1982, have stimulated public interest in the possibility of a volcano-climate interrelationship. Large volcanic dust veils in the atmosphere can reduce direct solar radiation by as much as 10% (Mass and Schneider 1977). This is simultaneously accompanied by an increased scattering effect which could substantially change the total amount of solar radiation reaching the earth's surface.

Many theoretical investigations (Schneider and Mass 1975; Pollack *et al.* 1976) and empirical studies (Lamb 1970; Oliver 1976; Mass and Schneider 1977; Taylor 1978) have been made in an attempt to determine the possible influence of large volcanic dust veils on surface weather and climate. Most of these investigations were conducted on either a global or hemispheric scale. Taylor (1978) also searched for volcanic signals on a latitudinal, continental/maritime and seasonal basis. A drop in average annual surface temperature of between 0.5 to 1.0°C in the first or second year following a large volcanic eruption was found in most of the empirical studies. The connection between large volcanic dust veils and average precipitation is less clear because none of the empirical investigations considered precipitation. Wexler (1951) suggested that volcanic dust particles might add to the supply of cloud condensation nuclei.

Canada, having an extensive area in mid-and high latitudes, should experience significant volcanic dust veil influences at varying times after a major eruption depending upon both the location and the time of year of the eruption. Oliver (1976) estimated a mid-latitude eruption to have a same-year impact on northern hemisphere mean temperatures, while a similar impact by an equatorial eruption would be delayed for about a year. Lamb (1970) states that the transfer of upper level dust veils from equatorial to mid-latitudes is accomplished mainly in autumn, and to a lesser extent in spring, with the great seasonal circulation changes.

---

[1] 1274 Parent Avenue, Windsor, Ontario, N8X 4J2. Term employee of Atmospheric Environment Service in 1983 as arranged through the Brock University internship program.

In this investigation, temperature and precipitation records for up to 20 Canadian stations were analyzed for evidence of volcanic dust veil signals. Investigations were made on national, regional (Arctic and Prairie) and seasonal (summer, winter, Prairie growing season) bases.[2]

METHODOLOGY AND DATA

The Superposed Epoch Method

The superposed epoch analysis method (Mass and Schneider 1977; Taylor 1978) was used in this investigation to detect changes in both Canadian surface temperatures and precipitation due to large volcanic dust veils. This method accentuates weak signals that are present in a data series. A temperature signal caused by a volcanic dust veil is expected to be of the same magnitude as the background noise level, or the variability of the atmosphere (Taylor 1978). The same relationship is expected between a possible precipitation signal and precipitation variability.

Volcanic Eruption Dates

Volcanic eruptions were selected on the basis of amounts of material injected into the atmosphere, latitude of the eruption and the isolation in time of the eruption from any other major volcanic event. Lamb's (1970) Dust Veil Index (DVI) was used to select most of the eruption dates. This is a measure of the amount of volcanic material injected into the atmosphere, and is directly related to the total loss of solar radiation reaching the surface of the earth. The eruption of Krakatau in 1883 was given a value of 1000, and all other eruptions were adjusted to it. The DVI has the advantage of not being calculated from temperature information (Mass and Schneider 1977).

Five of the six volcanic eruptions selected have a total DVI greater than 150, and are classed as major volcanic events (Table 1). The 1956 eruption was chosen because of its mid-latitude location and isolation from any other major volcanic event. Three equatorial and three mid-latitude eruptions were selected in an attempt to isolate both the temporal dimensions and the magnitudes in temperature and precipitation records of volcanic signals in Canada. Selected volcanic events were separated by at least five years from any other major event. This was done to avoid the problem of cumulative dust veils that might obscure resulting signals. The separation of the 1907 event from both the preceding event (1902) and the following (1912) is exactly five years. Incorporating the 1907 eruption was imperative as it was a major mid-latitude event for which ample records were available.

---

[2] Ed. note: See M. Parker's paper in this volume on tree rings, volcanic eruptions and climate in Canada.

Eruption-year key months were also determined from Lamb (1970). The key month was the month during which the volcano entered its most explosive phase. In the case of two or more eruptions in the same year, such as 1902, the month of the first eruption was used. Table 1 includes the key eruption date for each selected event.

Composite Key Dates

The key volcanic eruption date was defined as the 12-month period beginning with the month during which the eruption occurred. The use of this period results in a cleaner volcanic signal than using the actual calendar year of the eruption (Taylor 1978). The 12-month period was termed the "eruption year", or year "0". Sequences of four preceding years, or the four 12-month periods prior to the eruption year and four following years, or four 12-month periods after the eruption year, were then determined. These sequences provided the bases for both individual and multiple composites. The five annual periods, the eruption year and four following years, were analyzed because a volcanic dust veil produced by a single eruption exists for only a few years (Lamb 1970).

Composited Temperature and Precipitation

Average temperature and precipitation values, for selected Canadian stations, were calculated for each month of each of the 12-month periods associated with an eruption year. The resulting 12 monthly values were then summed and averaged to yield an annual value for that particular year. Graphs, based upon individual volcanic events were then plotted and studied in an attempt to define climatic signals.

Annual temperature and precipitation values for each individual eruption were then associated with the corresponding values for all other individual eruptions. Also, values

TABLE 1: *DATES, LOCATIONS AND DUST VEIL INDICES (DVI; SEE LAMB 1970) OF MAJOR VOLCANIC ERUPTIONS SELECTED FOR ANALYSIS.*

ERUPTIONS	LOCATIONS	DATES	KEY DATES	DVI	TOTAL DVI
1. Krakatau, Indonesia	6° S 105° E	August 1883	August 1883	1000	1000
2. Mont Pelée, Martinique	15° N 61° W	May 1902	- -	100	-
Soufrière, St. Vincent	13.5° N 61° W	May 1902	- -	300	-
Santa Maria, Guatemala	14.5° N 92° W	October 1902	- -	600	-
Cumulative data:			May 1902	-	1000
3. Shytubelya Sopka, Kamchatka	52° N 157.5° E	March 1907	March 1907	500	500
4. Katmai, Alaska	58° N 155° W	June 1912	June 1912	150	150
5. Bezymjannaja, Kamchatka	56° N 160.5° E	March 1956	March 1956	10	10
6. Gunung Agung, Bali	8.5° S 115.5° E	March 1963	March 1963	800	800

TABLE 2: *WEATHER STATIONS USED IN STUDYING THE INFLUENCE OF VOLCANIC DUST VEILS ON SURFACE TEMPERATURE AND PRECIPITATION RECORDS IN CANADA FOR THE 1883 KRAKATAU ERUPTION.*

WEATHER STATIONS	LOCATIONS	DATA PERIODS	YEARS OF RECORD	AES NUMBERS
1. Winnipeg, Manitoba	49° 53' N, 97° 07' W	1872-1938	67	5023243
2. Port Arthur, Ontario	48° 26' N, 89° 13' W	1877-1941	65	6046588
3. Ottawa, Ontario	45° 24' N, 75° 43' W	1872-1935	64	6105887
4. Beatrice, Ontario	45° 08' N, 79° 23' W	1876-1979	104	6110605
5. Woodstock, Ontario	43° 07' N, 80° 45' W	1870-1981	112	6149625
6. Toronto, Ontario	43° 40' N, 79° 24' W	1840-1981	142	6158350
7. Québec City, Québec	46° 48' N, 71° 13' W	1872-1959	88	7016280
8. Montréal, Québec	45° 30' N, 73° 35' W	1871-1981	111	7025280
9. Chatham, New Brunswick	47° 03' N, 65° 29' W	1873-1947	75	8100990
10. Fredericton, New Brunswick	45° 57' N, 66° 36' W	1871-1952	82	8101700
11. Halifax, Nova Scotia	44° 39' N, 63° 36' W	1871-1933	63	8202198
12. Sydney, Nova Scotia	46° 09' N, 60° 12' W	1870-1941	72	8205698
13. St. John's, Newfoundland	47° 34' N, 52° 42' W	1874-1956	83	8403500

for equatorial eruptions were isolated and inter-associated. The same was done with mid-latitude eruption values. These corresponding values were then summed and averaged to yield a "superposed epoch". In other words, for all the data, or one of the subsets mentioned, all eruption year values, all values of eruption year plus one, etc. were summed and averaged. Graphs, based upon these multiple volcanic events, were plotted and studied in a manner comparable to analyses of the individual events.

Data

The database used in this investigation consisted of mean monthly temperature values and total monthly precipitation values. Up to 20 different Canadian weather stations were used over common time periods.

Stations were selected on the basis of length and completeness of record and upon location. Thirteen stations were available for the 1883 eruption date. There were no long-term records available for this date west of Winnipeg. Four stations were added for the 1902 eruption date to provide an east to west coast coverage. Another station was added for the 1907 and 1912 eruptions. The lack of long-term stations in northern regions restricts the study of the first four eruptions to more southerly Canadian latitudes. Northern stations were added for the last two eruptions, bringing the total to 20 stations for the 1963 event. Table 2 shows the stations used for the 1883 eruption. Table 3 shows the stations added for the 1902, 1907 and 1912 eruptions. Table 4 lists the stations used for the 1956 and 1963 eruptions. In some cases, such as Québec City and Winnipeg, weather observation sites were moved during the 1940s from city to airport. None of the eruptions used in this study occurred during this period.

Missing monthly values were estimated for each station by calculating the 30-year mean for that particular month. In most cases only one of the 13 to 20 values was absent. The resulting estimate had little effect on the overall monthly composite. There were never more than two missing values in any monthly composite.

TABLE 3: *STATIONS ADDED TO THOSE USED FOR THE 1883 KRAKATAU ERUPTION FOR THE 1902, 1907, 1912 ERUPTIONS.*

	WEATHER STATIONS[1]	LOCATIONS	DATA PERIODS	YEARS OF RECORD	AES NUMBERS
1.	Victoria, British Columbia	48° 25' N, 123° 22' W	1898-1981	84	1018610
2.	Medicine Hat, Alberta	50° 01' N, 110° 43' W	1883-1981	99	3034480
3.	Banff, Alberta	51° 11' N, 115° 34' W	1887-1981	95	3050520
4.	Regina, Saskatchewan	50° 26' N, 104° 40' W	1883-1981	99	4016560
5.	Ottawa (CDA), Ontario	45° 23' N, 75° 43' W	1889-1981	93	6105976

[1] Ottawa (6105887) not used for 1902 eruption.

TABLE 4: *WEATHER STATIONS USED IN STUDYING THE INFLUENCE OF VOLCANIC DUST VEILS ON SURFACE TEMPERATURE AND PRECIPITATION RECORDS IN CANADA FOR THE 1956 BEZYMJANNAJA AND 1963 GUNUNG AGUNG ERUPTIONS.*

	WEATHER STATIONS[1]	LOCATIONS	DATA PERIODS	YEARS OF RECORD	AES NUMBERS
1.	Victoria, British Columbia	48° 25' N, 123° 22' W	1898-1981	84	1018610
2.	Medicine Hat, Alberta	50° 01' N, 110° 43' W	1883-1981	99	3034480
3.	Banff, Alberta	51° 11' N, 115° 34' W	1887-1981	95	3050520
4.	Regina, Saskatchewan	50° 26' N, 104° 40' W	1883-1981	99	4016560
5.	Winnipeg (A), Manitoba	49° 54' N, 97° 14' W	1938-1981	44	5023222
6.	Churchill, Manitoba	58° 45' N, 94° 04' W	1943-1981	39	5060600
7.	Ottawa (CDA), Ontario	45° 23' N, 75° 43' W	1889-1981	93	6105976
8.	Beatrice, Ontario	45° 08' N, 79° 23' W	1876-1979	104	6110605
9.	Woodstock, Ontario	43° 07' N, 80° 45' W	1870-1981	112	6149625
10.	Toronto, Ontario	43° 40' N, 79° 24' W	1840-1981	142	6158350
11.	Québec City (A), Québec	46° 48' N, 71° 23' W	1943-1981	39	7016294
12.	Montréal, Québec	45° 30' N, 73° 35' W	1871-1981	111	7025280
13.	Chatham (A), New Brunswick	47° 01' N, 65° 27' W	1943-1981	39	8101000
14.	Fredericton (CDA), New Brunswick	45° 55' N, 66° 37' W	1913-1981	69	8101600
15.	Halifax, Nova Scotia	44° 39' N, 63° 34' W	1939-1974	36	8202200
16.	Sydney (A), Nova Scotia	46° 10' N, 60° 03' W	1941-1981	41	8205700
17.	St. John's, Newfoundland	47° 35' N, 52° 44' W	1957-1975	19	8403501
18.	Cambridge Bay, NWT	69° 07' N, 105° 01' W	1929-1981	53	2500600
19.	Mould Bay, NWT	76° 14' N, 119° 20' W	1948-1981	34	2502700
20.	Fort Chimo, Québec	58° 06' N, 68° 25' W	1947-1981	35	7112400

[1] St. John's (8403501) not used for 1956 eruption.

CANADIAN ANALYSIS

Taylor (1978) found it necessary to use data from a group of stations rather than just individual stations when searching for temperature signal related to a volcanic eruption. This is due to the year-to-year and station-to-station variability when dealing with single station superpositions. Thus, the superposed epoch analysis method outlined previously was applied to the entire temperature and precipitation database selected for this study (maximum of 20 stations).

Temperature

Individual Eruption Composites

Figures 1-6 show the individual eruption dust veil temperature composites for selected Canadian stations. The 1883, 1902, 1956 and 1963 composites each display a marked dip in average annual temperature either in the eruption year or in the following two years. The 1907 and 1912 composites show no such dip during these years. The low values for the "-4" and "-3" years for the 1907 composite might be the result of a large 1902 dust veil, but, there is no such dip in the early years of the 1912 composite.

The apparent significance of these graphs must be viewed with caution. The first four eruptions were imbedded in a hemispheric warming trend, whereas the last two eruptions occurred during a hemispheric cooling trend (Mass and Schneider 1977). The year-to-year variability found by Taylor (1978) is evident in a Canadian context. The compositing of several volcanic events should reduce this variability.

Multiple Eruption Composites

Figure 7 shows the temperature composite for all stations and all eruptions. The year-to-year variability has been substantially reduced. *There is an obvious temperature dip during the eruption year and in the "+1" year.* The dip during these two years is about 0.4°C below the level of years "-4" to "-1". Figure 8 shows the composite for the three equatorial eruptions. There is a well-marked dip in the "+1" year, about 1.1°C below the level of years "-4" to "-1". Figure 9 shows the composite for the three mid-latitude

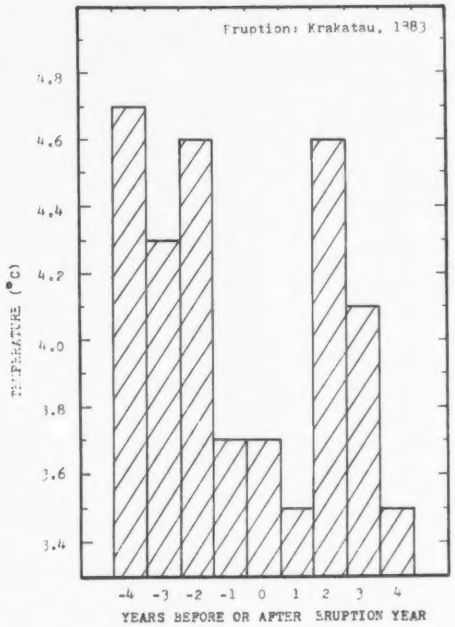


FIGURE 1: *Dust veil temperature composite. Thirteen Canadian station events.*

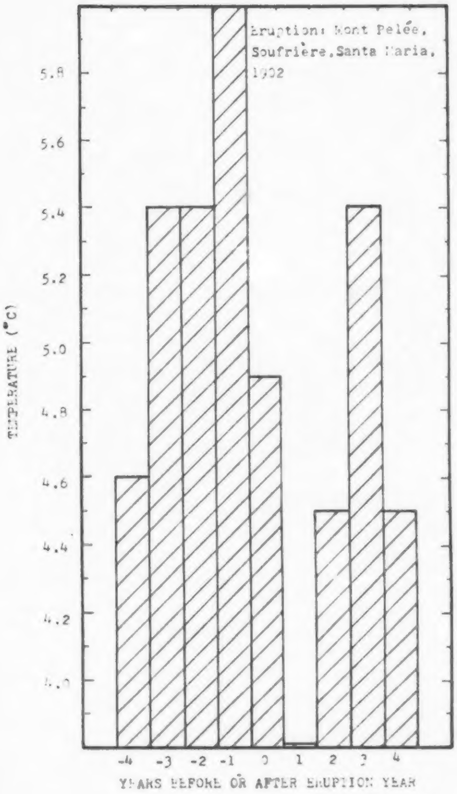


FIGURE 2: *Dust veil temperature composite. Seventeen Canadian station events.*

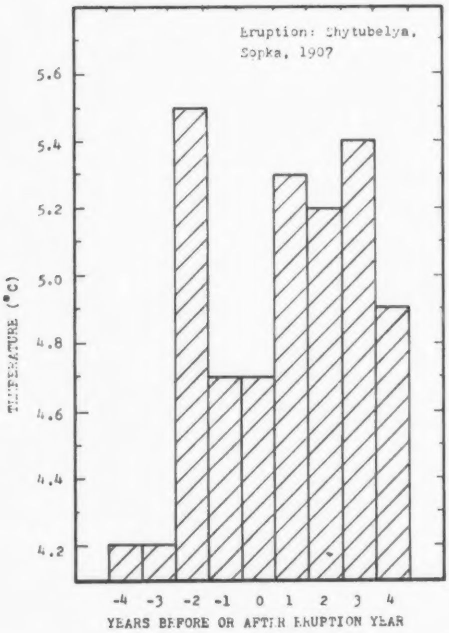


FIGURE 3: *Dust veil temperature composite. Eighteen Canadian station events.*

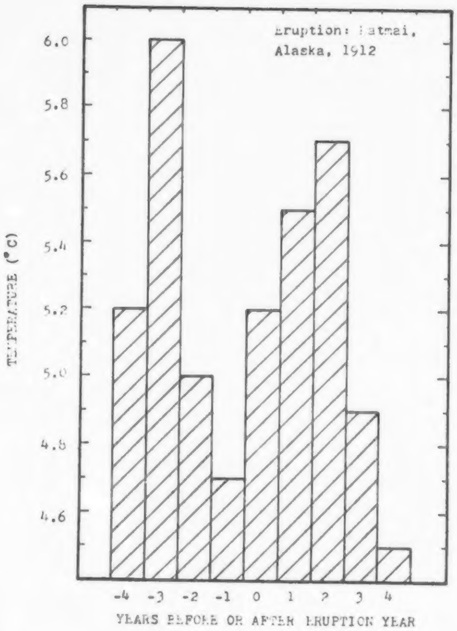


FIGURE 4: *Dust veil temperature composite. Eighteen Canadian station events.*

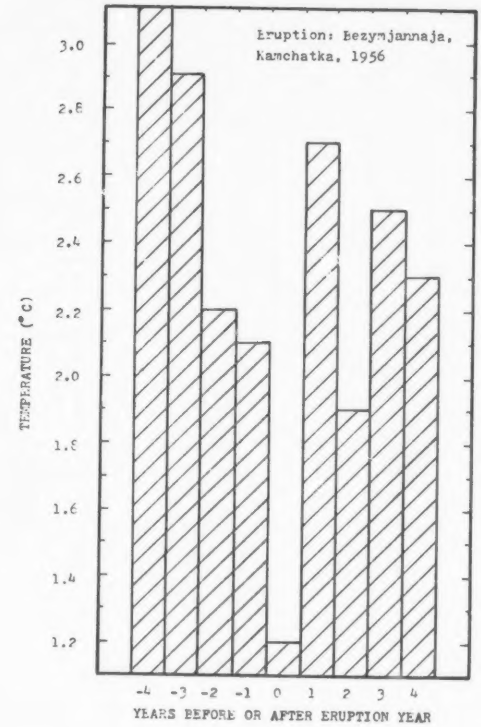


FIGURE 5: *Dust veil temperature composite. Nineteen Canadian station events.*

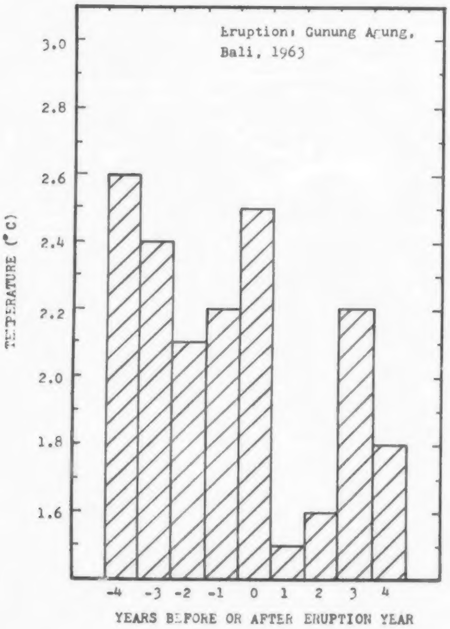


FIGURE 6: *Dust veil temperature composite. Twenty Canadian station events.*

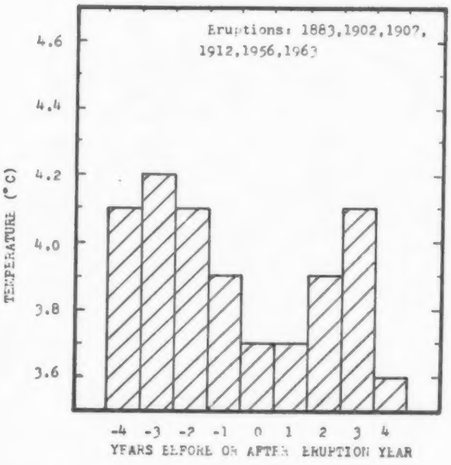


FIGURE 7: *Dust veil temperature composite. All Canadian station events.*

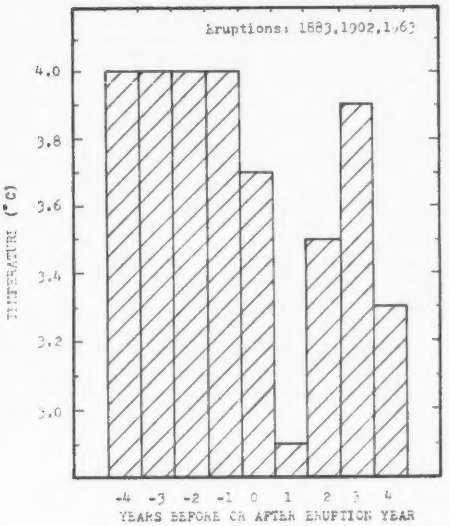


FIGURE 8: *Dust veil temperature composite. Fifty Canadian station events for three equatorial eruptions.*

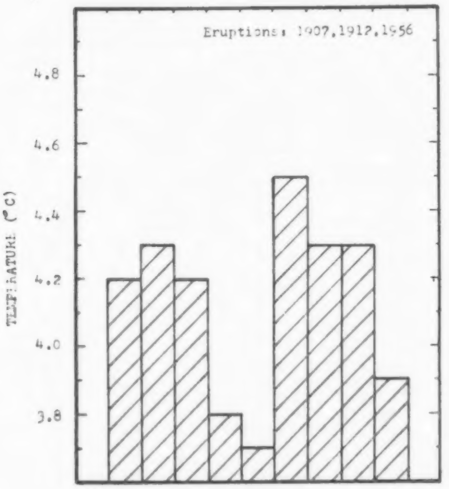


FIGURE 9: *Dust veil temperature composite. Fifty-four Canadian station events for three mid-latitude eruptions.*

eruptions. Here the temperature dip is in the actual eruption year, about 0.5°C below the level of years "-4" to "-1".

Precipitation

Individual Eruption Composites

Figures 10-15 show the individual eruption dust veil precipitation composites for selected Canadian stations. There is the same year-to-year variability which was evident in the individual temperature composites. The 1902 and 1912 eruptions both display marked drops in average annual total precipitation during the "+1", "+2" or "+3" years. No such drop is evident in any of the other composites.

Multiple Eruption Composites

Figure 16 shows the precipitation composite for all stations and all eruptions. The year-to-year variability, as with temperature, has been greatly reduced. There is a dip in the "+1" and "+2" years about 3mm below the level of years "-4" to "-1". Figure 17 shows the composite for the three equatorial eruptions. The dip in years "+2" and "+3" is about 5 mm below the level of years "-4" to "-1". Figure 18 shows the composite for the three mid-latitude eruptions. The slight dip in the year following the eruption year is about 3.5 mm below the level of years "-4" to "-1".

REGIONAL AND SEASONAL ANALYSES

The data noise level (year-to-year variability), when based upon different groupings of stations, should vary randomly while the volcanic signal should remain fairly constant (Taylor 1978). A regional analysis was one step in determining the significance of the possible volcanic signals outlined previously. It also provided the basis for volcanic signal investigations in parts of Canada which might be sensitive to small alterations in temperature or precipitation, or both. A summer/winter seasonal investigation was also made.

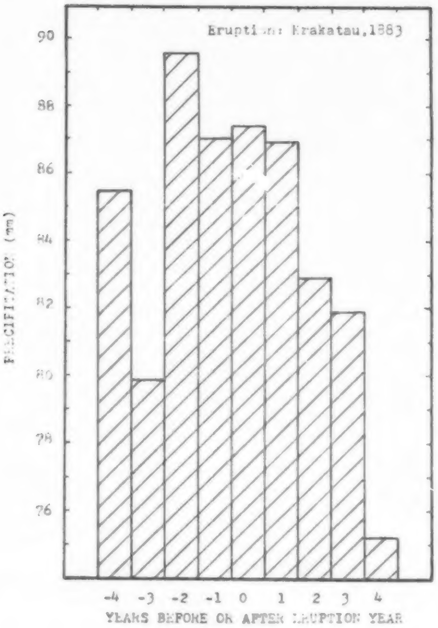


FIGURE 10: *Dust veil precipitation composite. Thirteen Canadian station events.*

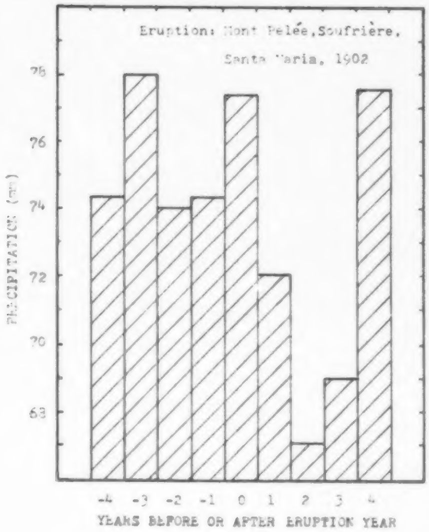


FIGURE 11: *Dust veil precipitation composite. Seventeen Canadian station events.*

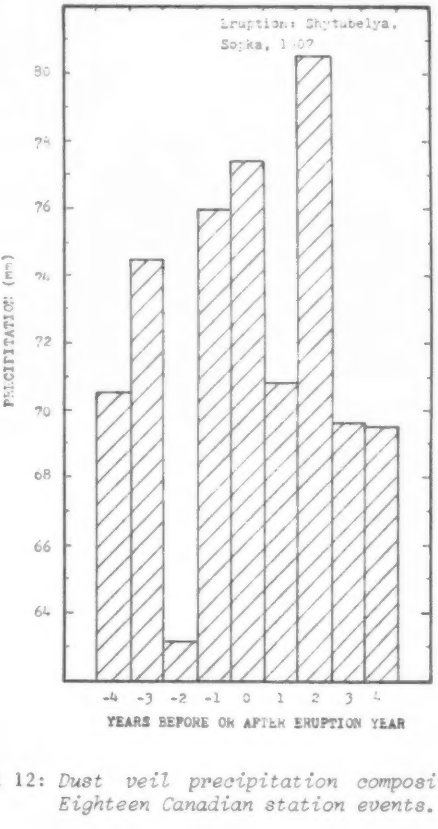


FIGURE 12: *Dust veil precipitation composite. Eighteen Canadian station events.*

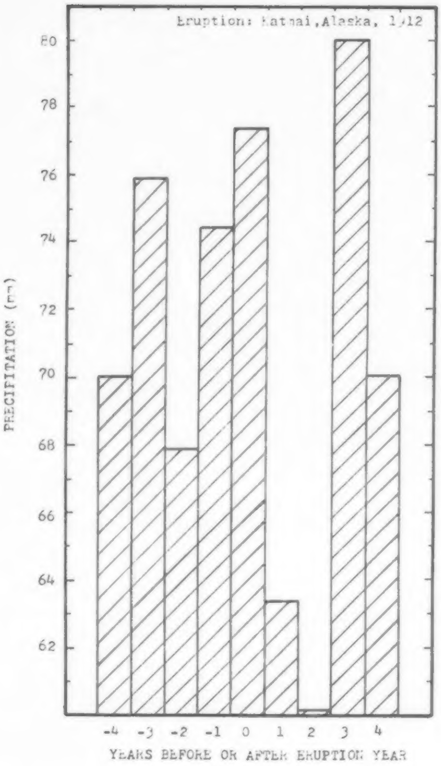


FIGURE 13: *Dust veil precipitation composite. Eighteen Canadian station events.*

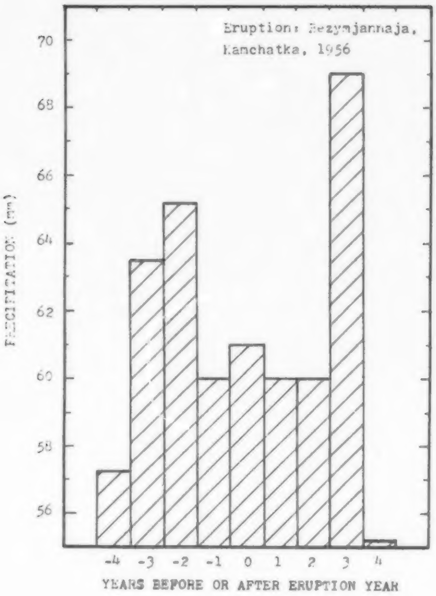


FIGURE 14: *Dust veil precipitation composite. Nineteen Canadian station events.*

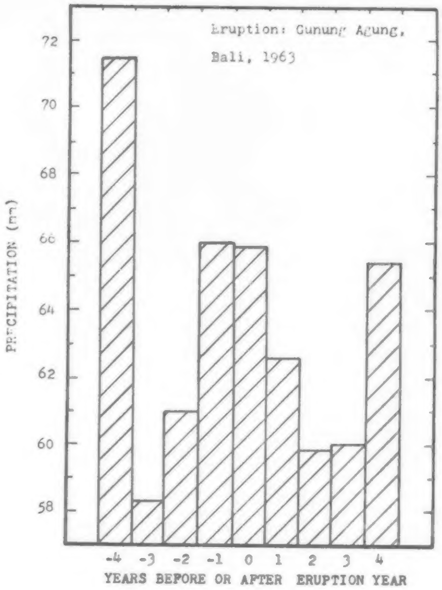


FIGURE 15: *Dust veil precipitation composite. Twenty Canadian station events.*

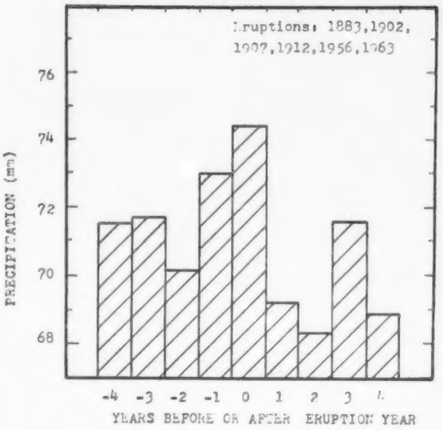


FIGURE 16: *Dust veil precipitation composite. All Canadian station events.*

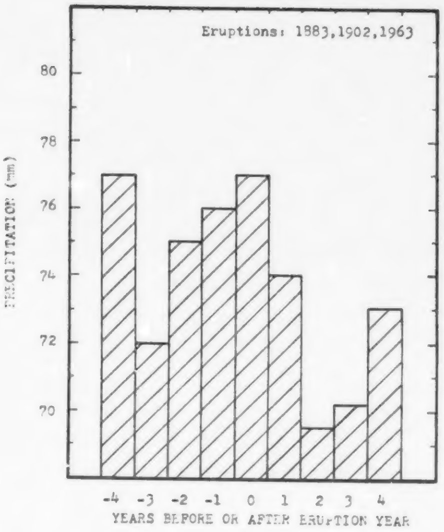


FIGURE 17: *Dust veil precipitation composite. Fifty Canadian station events for three equatorial eruptions.*

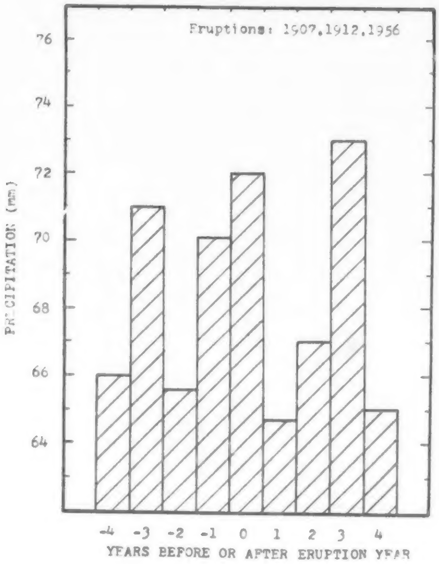


FIGURE 18: *Dust veil precipitation composite. Fifty-four Canadian station events for three mid-latitude eruptions.*

Regional

Arctic Analysis

The solar radiation deficit produced by volcanic dust veils must be greatest in Arctic areas where dust veils persist longest and the sun's rays travel obliquely through the layers of dust (Lamb 1970). Reduced surface temperatures result in an accumulation of both sea ice and land snow. The increased albedo would produce a radiation deficit long after the dust veil has disappeared (Lamb 1970). It would also affect the general atmospheric circulation, possibly having far-reaching spatial effects.

The superposed epoch analysis method was applied to four Canadian Arctic stations for the 1956 and 1963 eruptions. There were no Arctic station records for earlier eruptions. The stations used were Churchill, Manitoba, Cambridge Bay, Northwest Territories, Mould Bay, Northwest Territories and Fort Chimo, Québec.

Figure 19 shows the average annual temperature composite while Figure 20 shows the average annual total precipitation composite for the 1956 mid-latitude and 1963 equatorial eruptions. The temperature dip in the "0" and "+1" years is similar to that in Figure 7 for all Canadian stations. It is about 1°C below the level of years "-4" to "-1". The surrounding noise level, however, is quite different from that in Figure 7. Years "+2" to "+4" hint at Arctic temperature stability following a volcanic eruption. The precipitation dip, beginning possibly in year "+1" and culminating in year "+2" is similar to that in Figure 16 for all stations. It is about 2.5 mm below the level of years "-4" to "-1". The surrounding noise levels for the two graphs are quite different.

Prairie Analysis

A Prairie region precipitation signal was investigated in light of the possible signal for the entire country oulined previously. Four to six stations were selected for five eruptions beginning with 1902. Table 3 shows the stations used for each eruption. Figures 21-23 show the average annual total precipitation composites for these eruptions. The drop in years "+1" and "+2" for all eruptions in Figure 21 is similar to that in Figure 16 for all stations. It is about 3.5 mm below the level of years "-4" to "-1". The stepped pattern in years "+2" to "+4" could possibly be a recovery period to normal precipitation levels. The only signal apparent in Figure 22, for equatorial eruptions, is the

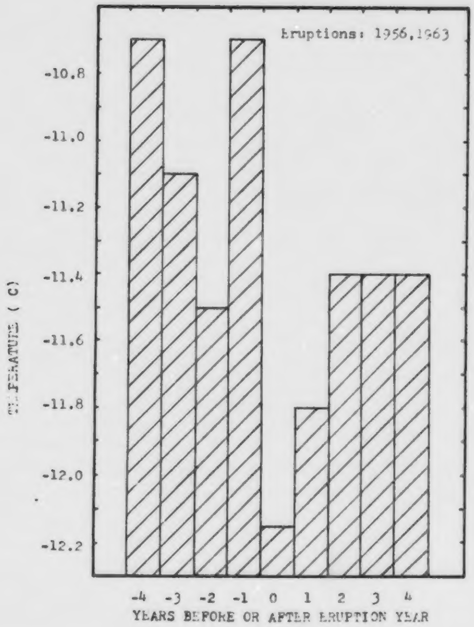


FIGURE 19: *Dust veil temperature composite. Canadian Arctic station events.*

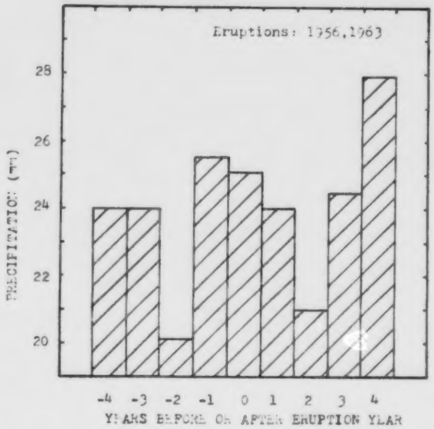


FIGURE 20: *Dust veil precipitation composite. Canadian Arctic station events.*

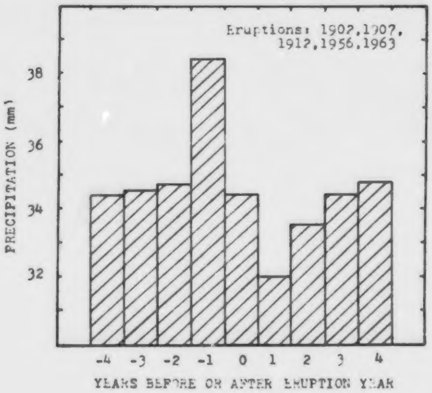


FIGURE 21: *Dust veil precipitation composite. Canadian Prairie station events.*

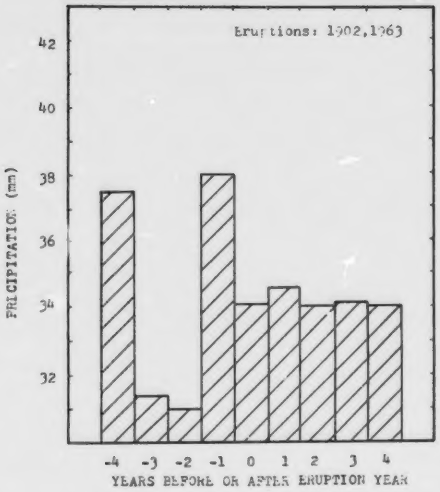


FIGURE 22: *Dust veil precipitation composite. Canadian Prairie station events.*

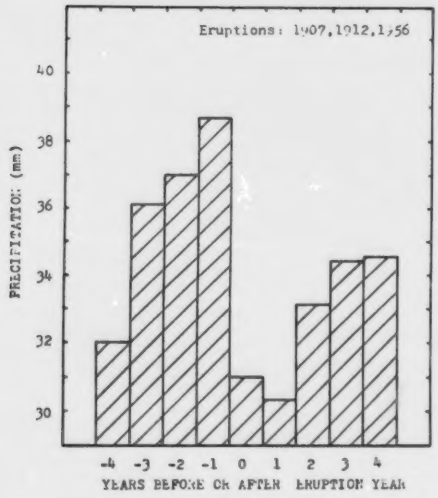


FIGURE 23: *Dust veil precipitation composite. Canadian Prairie station events.*

year-to-year stability beginning with the eruption year and continuing to year "+4". Figure 23, the mid-latitude composite, shows a distinct dip during both the eruption year and year "+1". It is much more of a decrease in precipitation than that for all Canadian stations in Figure 18. The dip is about 5 mm below the level of years "-4" to "-1". The surrounding noise levels, for each graph, are different from those for all stations.

Seasonal Analyses

Summer and winter investigations were made in an attempt to determine the relative magnitudes of the dust signals. The key summer season was defined as the first three-month period (June-August), whereas the key winter season was defined as the first three-month period (December-February) to follow an eruption. These seasons were termed the "0" or eruption year. Sequences of four preceding and four following seasons were determined in the same manner outlined previously. Seasonal averages were calculated for all Canadian stations and for each year associated with a volcanic eruption. Graphs, based upon multiple volcanic events were then plotted and examined in an attempt to define volcanic signals. Finally, the Prairie stations were used in an attempt to detect a precipitation signal during the May-September growing season.

Summer

Figures 24-29 show the temperature and precipitation summer season composites. All but one graph show a distinct drop of up to several tenths of a degree or a few millimetres in either the eruption year or the following year. These composites display a close resemblance to those in Figures 7-9 for annual temperature and Figures 16-18 for annual precipitation. The magnitude of each drop, however, is at least equal to or greater than that in the corresponding annual composite.

Winter

Figures 30-35 show the temperature and precipitation winter season composites. There is a higher degree of year to year variability than was evident in the summer composites. This makes it more difficult to detect a volcanic signal. There is a distinct drop in temperature

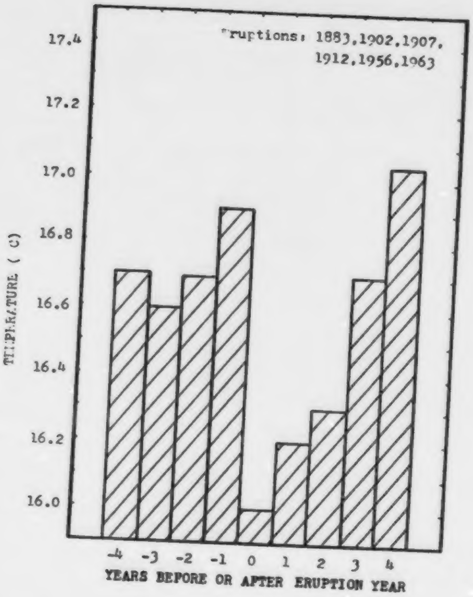


FIGURE 24: *Dust veil summer season temperature. All Canadian stations.*

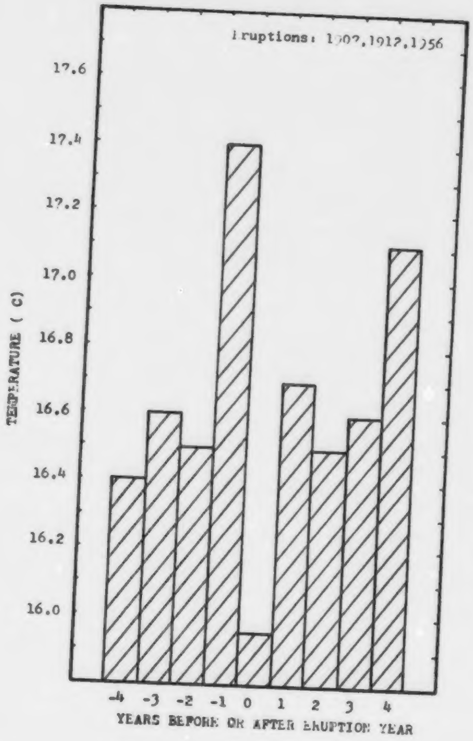


FIGURE 26: *Dust veil summer season temperature composite. All Canadian stations.*

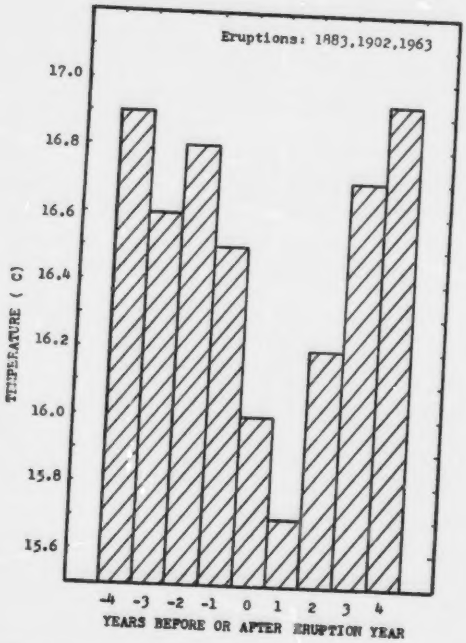


FIGURE 25: *Dust veil summer season temperature composite. All Canadian stations.*

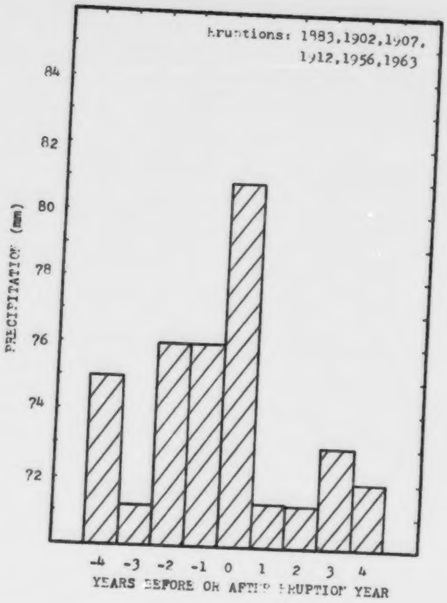


FIGURE 27: *Dust veil summer season precipitation composite. All Canadian stations.*

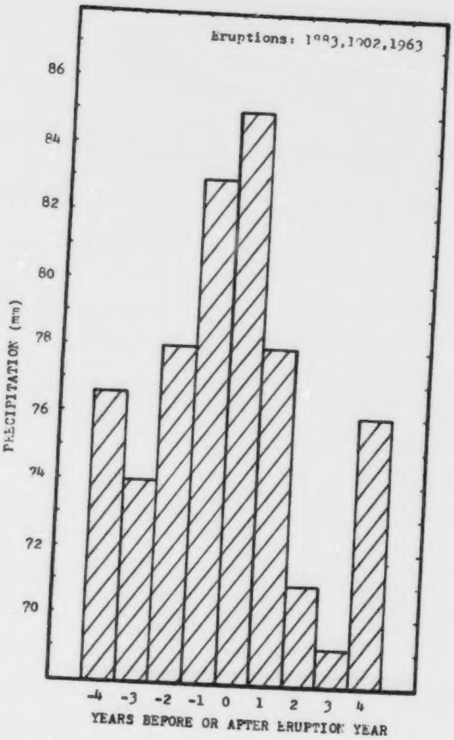


FIGURE 28: *Dust veil summer season precipitation composite. All Canadian stations.*

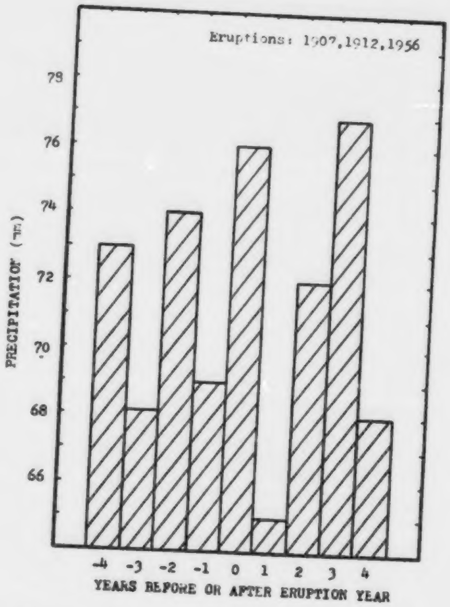


FIGURE 29: *Dust veil summer season precipitation composite. All Canadian stations.*

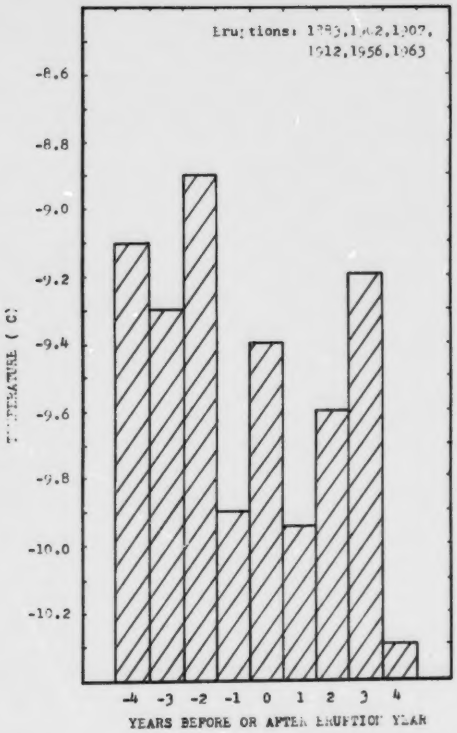


FIGURE 30: *Dust veil winter season temperature composite. All Canadian stations.*

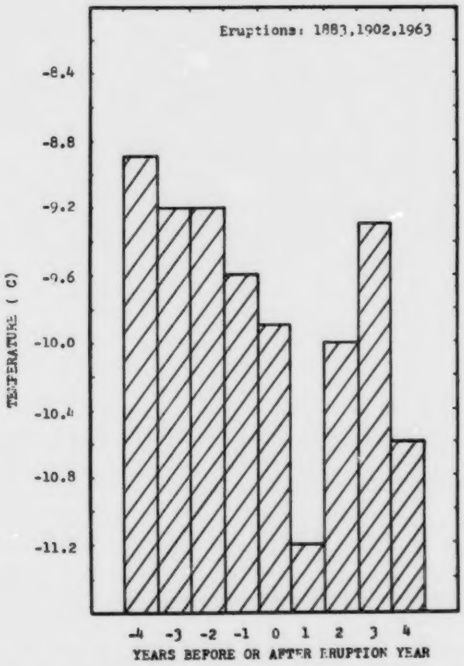


FIGURE 31: *Dust veil winter season temperature composite. All Canadian stations.*

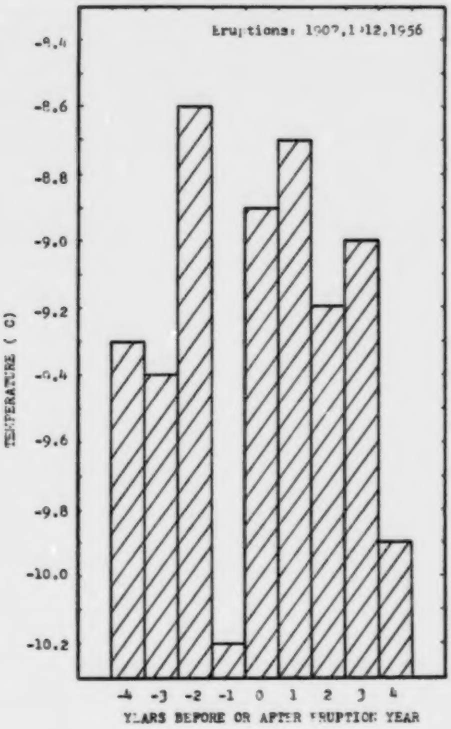


FIGURE 32: *Dust veil winter season temperature composite. All Canadian stations.*

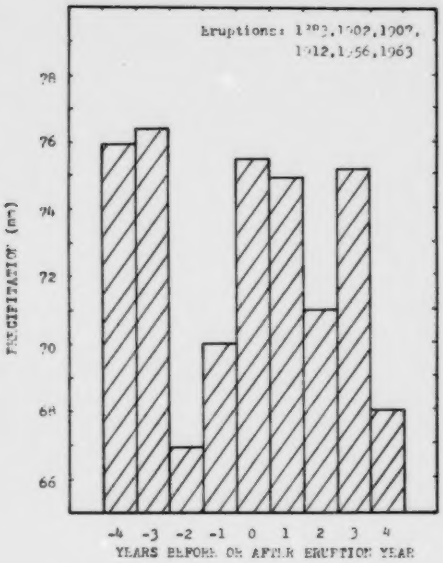


FIGURE 33: *Dust veil winter season precipitation composite. All Canadian stations.*

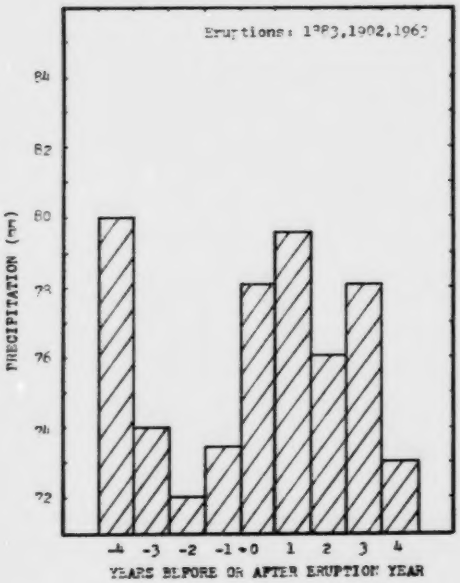


FIGURE 34: *Dust veil winter season precipitation composite. All Canadian stations.*

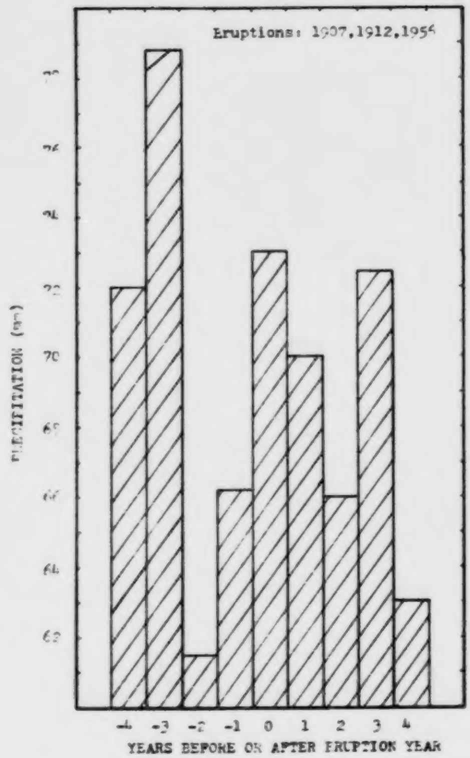


FIGURE 35: *Dust veil winter season precipitation composite. All Canadian stations.*

during the first winter following an equatorial eruption (Figure 31) but there is no such drop following a mid-latitude eruption (Figure 32). No precipitation signal is evident after an equatorial (Figure 34) or a mid-latitude (Figure 35) eruption. In general, these graphs resemble their annual counterparts in Figures 7-9 and 16-18 but the higher degree of year-to-year variability completely obscures volcanic signals in most cases.

Prairie Growing Season Precipitation

The key Prairie growing season was defined arbitrarily as the five-month period (May-September) following a volcanic eruption. Figures 36-38 show the growing season precipitation composites for the stations in Tables 5 and 6. These graphs resemble those in Figures 21-23. No signal is evident following an equatorial eruption (Figure 37). There is a distinct drop in the "+1" year following a mid-latitude eruption (Figure 38). It is about 13 mm below the level of years "-4" to"-1". The stepped pattern of increase in years "+2" and "+3" is also evident as it was in Figure 23. The signal in Figure 36 for all eruptions can be attributed primarily to the mid-latitude events.

SIGNIFICANCE TESTS

The fact that the regional Arctic study identified much the same volcanic signals as those of the national study is a supportive indication of significance. A more rigorous small sample test, however, is desirable.

The Student t-test was applied to the multiple eruption composites to determine whether the sample mean, or the mean of the one or two years during which the volcanic signal is evident, is significantly different from the population mean, or the mean of the nine years from which it was taken. Mass and Schneider (1977) applied this test to volcanic dust veil composites for northern hemisphere stations. The basic formula used was:

TABLE 5: *WEATHER STATIONS USED IN STUDYING THE INFLUENCE OF VOLCANIC DUST VEILS ON PRECIPITATION RECORDS IN THE CANADIAN PRAIRIES FOR THE 1902, 1907 SHYTUBELYA AND 1912 KATMAI ERUPTIONS.*

	WEATHER STATIONS	LOCATIONS		DATA PERIODS	YEARS OF RECORD	AES NUMBERS
1.	Medicine Hat, Alberta	50° 01' N,	110° 43' W	1883-1981	99	3034480
2.	Regina, Saskatchewan	50° 26' N,	104° 40' W	1883-1981	99	4016560
3.	Swift Current, Saskatchewan	50° 12' N,	107° 48' W	1885-1938	54	4028035
4.	Winnipeg, Manitoba	49° 53' N,	97° 07' W	1872-1938	67	5023243

TABLE 6: *WEATHER STATIONS USED IN STUDYING THE INFLUENCE OF VOLCANIC DUST VEILS ON PRECIPITATION RECORDS IN THE CANADIAN PRAIRIES FOR THE 1956 BEZYMJANNAJA AND 1963 GUNUNG AGUNG ERUPTIONS.*

	WEATHER STATIONS	LOCATIONS		DATA PERIODS	YEARS OF RECORD	AES NUMBERS
1.	Medicine Hat, Alberta	50° 01' N,	110° 43' W	1883-1981	99	3034480
2.	Lethbridge, Alberta	49° 40' N,	112° 50' W	1936-1981	46	3033880
3.	Regina, Saskatchewan	50° 26' N,	104° 40' W	1883-1981	99	4016560
4.	Swift Current (A), Saskatchewan	50° 12' N,	107° 48' W	1938-1981	44	4028040
5.	Moose Jaw (A)[1], Saskatchewan	50° 17' N,	105° 35' W	1938-1981	44	4015320
6.	Winnipeg (A), Manitoba	49° 54' N,	97° 14' W	1938-1981	44	5023222

[1] Moose Jaw used only for 1963 eruption.

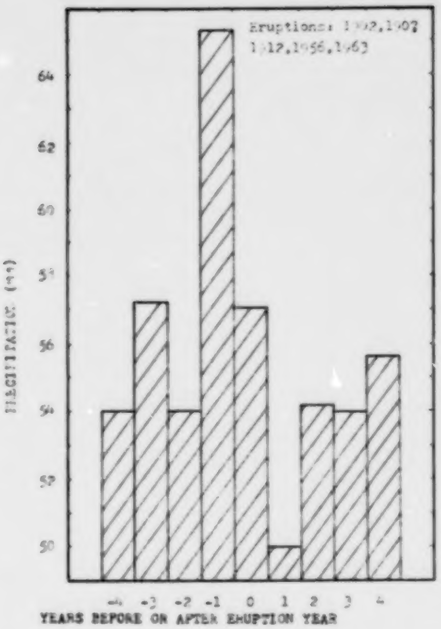


FIGURE 36: *Dust veil growing season (May-Sept.) precipitation composite. Canadian Prairie stations.*

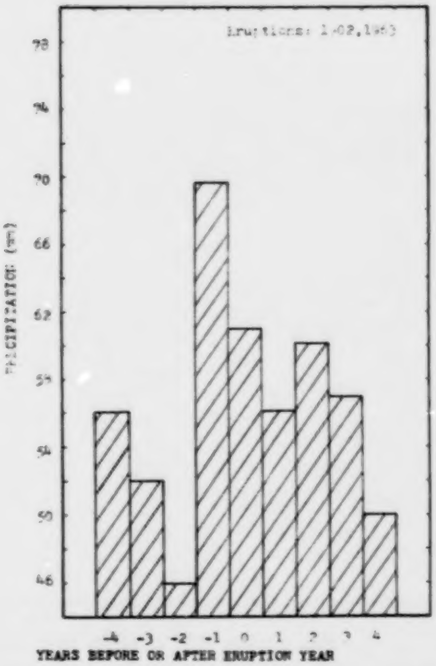


FIGURE 37: *Dust veil growing season (May-Sept.) precipitation composite. Canadian Prairie stations.*

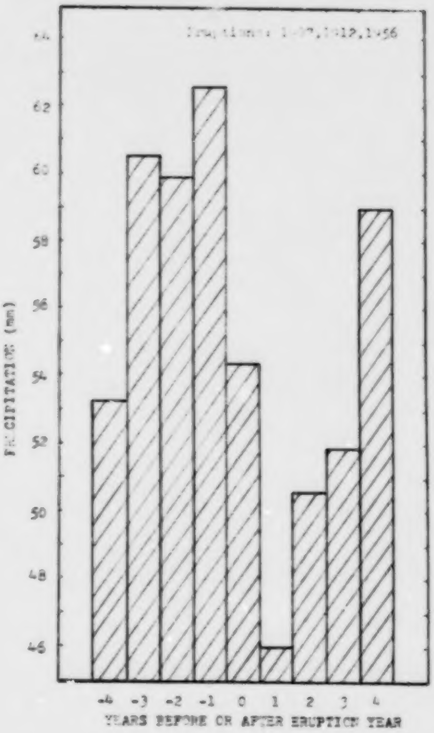


FIGURE 38: *Dust veil growing season (May-Sept.) precipitation composite. Canadian Prairie stations.*

[1] $$t = \frac{[\bar{x} - \mu]\sqrt{N}}{\sigma}$$

where, $\bar{x}$ = sample mean
μ = population mean
σ = population standard deviation

and, N = average number of stations x number of eruptions in composite

Thus, for a mid-latitude composite for all stations

$$N = 18 \times 3 = 54$$

or, for an all-eruption Prairie composite

$$N = 5 \times 5 = 25$$

The degrees of freedom (Gregory 1963) are:

$$\text{d.f.} = (n - 1) + (n - 1)$$
$$= n_a + n_b - 2$$

where, n_a = number of population years
n_b = number of sample years
so, d.f. (1 sample year) = 8
d.f. (2 sample years) = 9

The problem dealt with earlier concerning the low number of stations and eruptions when dealing with Arctic and Prairie composites needs to be borne in mind. The fewer stations or eruptions used, the greater the difference between the means must be, in order to reach a given level of significance. Table 7 shows the calculated Student-t values and the associated levels of significance for all composites where a volcanic signal was apparent. The test results are similar to those of Mass and Schneider (1977) (i.e. there is a difference between the sample population and the entire set at a significance level (α) of at least 99%).

CONCLUSIONS

Variation in climate is complex and influenced by many factors. Clear identification of possible volcanic influences is therefore difficult. Consequently caution must be used in interpreting apparent historic evidence, and using it to predict future events. Nevertheless, the results of this study do provide some evidence of effects of volcanic influences -particularly on temperatures- and permit some tentative conclusions to be made.

The magnitude of the annual temperature drop, for all Canadian stations, was at least 0.5°C greater after the equatorial eruptions analyzed than after the mid-latitude eruptions. The average total DVI for the selected equatorial eruptions was 700 while it was

TABLE 7: *STUDENT t-TEST CALCULATIONS FOR COMPOSITES HAVING AN APPARENT VOLCANIC DUST VEIL SIGNAL.*

		FUNCTIONS USED IN SIGNIFICANCE TESTS							
FIGURES	COMPOSITE	$\bar{x}$	YEARS	μ	σ	N	d.f.	t	α (%)
Fig. 7	all stn. all events (temperature)	3.70	(0,1)	3.91	0.21	108	9	10.4	99.9
Fig 8	all stn. eq. events (temperature)	2.91	(1)	3.70	0.37	51	8	15.2	99.9
Fig. 9	all stn. mid-lat. events (temperature)	3.68	(0)	4.13	0.25	54	8	13.2	99.9
Fig. 16	all stn. all events (precipitation)	68.77	(1,2)	70.96	1.89	108	9	12.0	99.9
Fig. 17	all stn. eq. events (precipitation)	69.53	(2)	73.67	2.58	51	8	11.5	99.9
Fig. 18	all stn. mid-lat. events (precipitation)	64.65	(1)	68.24	2.87	54	8	9.2	99.9
Fig. 19	Arctic temperature (a)	-11.97	(0,1)	-11.34	0.44	8	9	4.05	99.0
	(b)	-12.15	(0)				8	5.2	99.9
Fig. 20	Arctic precipitation	20.93	(2)	24.00	2.15	8	8	4.0	99.0
Fig. 21	Prairie, all events (precipitation)	31.96	(1)	34.48	1.58	25	8	8.0	99.9
Fig. 23	Prairie, mid-lat. events (precipitation) (a)	30.67	(0,1)	34.17	2.65	12	9	4.6	99.0
	(b)	30.29	(1)				8	5.1	99.9
Fig. 24	all stn. all events (summer temp.)	16.09	(0,1)	16.57	0.33	108	9	15.1	99.9
Fig. 25	all stn. eq. events (summer temp.)	15.93	(1,2)	16.48	0.41	51	9	9.6	99.9
Fig. 26	all stn. mid-lat. events (summer temp.)	15.95	(0)	16.64	0.39	54	8	13.0	99.9
Fig. 27	all stn. all events (summer precip.)	71.35	(1,2)	74.02	2.97	108	9	9.3	99.9
Fig. 28	all stn. eq. events (summer precip.)	70.19	(2,3)	76.78	4.80	51	9	9.8	99.9
Fig. 29	all stn. mid-lat. events (summer precip.)	64.90	(1)	71.26	3.73	54	8	12.5	99.9
Fig. 31	all stn. eq. events (winter temp.)	-10.63	(1,2)	-9.76	0.72	51	9	8.6	99.9
Fig. 38	Prairie grow. season mid-lat. events (a)	48.22	(1,2)	55.33	5.21	12	9	4.7	99.0
	(precipitation) (b)	45.96	(1)				8	6.2	99.9

220 for the mid-latitude events. The mid-latitude temperature depression was about 0.4°C, occurring during the eruption year and lasting no longer. The equatorial signal of about 1°C occurred during the first year after the eruption year and persisted to a lesser degree, for another year or so.

The time and magnitude of an annual precipitation drop, for all Canadian stations, is also related to the latitude of the eruption and the size of the dust veil respectively. A 5-mm drop was evident during the second and third years after the equatorial eruptions. A 3.5-mm drop occurred in the first year after the mid-latitude eruptions.

The annual temperature drop in the Arctic was slightly greater than that for the country as a whole. It was about 1°C, and occurred in both the eruption year and the year following. There appears to have been a small Arctic precipitation signal of about 2.5 mm in the second year after an eruption year. The lower significance levels for the Arctic signals reflects the small number of stations and events used. Further investigation of this region, using more stations, might be appropriate.

The decline in annual Prairie precipitation was greatest after a mid-latitude eruption. A drop of about 5 mm occurred during the eruption year and the following year.

Both temperature and precipitation signals were stronger in the summer than in the winter. Summer drops in temperature and precipitation were, in almost all cases, of a greater magnitude than the annual drops. There was a marked drop in winter temperature of about 1°C in the year following an equatorial eruption.

There was a greater drop in Prairie precipitation during the (May-September) growing season and after the mid-latitude eruptions than on an annual basis. This was an extra 8 mm, or about 13 mm. Further investigation into the relationship between Prairie precipitation and volcanic eruptions is needed. The study should have an expanded database with more emphasis on mid-latitude eruptions and incorporate precipitation trends in the region.

This investigation did not take temperature or precipitation *trends* into account. No technique, other than the compositing of several volcanic events, was used to eliminate trends. The first four eruptions selected occurred during hemispheric warming trends, whereas the latter two occurred during hemispheric cooling trends. An accurate assessment of volcanic dust veil signals would eliminate these trends before applying the compositing technique. Precipitation trends, being more regional in nature than temperature trends, could pose a greater problem.

The temperature results found in this investigation are quantitatively similar to the empirical results found by Mass and Schneider (1977) and Taylor (1978) and to the theoretical results found by Pollack *et al.* (1976). There are no comparable precipitation results. The reduction in precipitation in Canada, on all levels, was quite different from the increase suggested by the condensation nuclei hypothesis (Wexler 1951).

SUMMARY

Temperature and precipitation data from 20 Canadian weather stations were merged with Lamb's dust veil indices for major volcanic eruptions since 1883 using a compositing technique (Mass and Schneider 1977, Taylor 1978). The method enhances the volcanic signal, making it possible to detect a weak temperature response. A less conclusive decrease in overall precipitation following the studied volcanic events is also indicated.

ACKNOWLEDGEMENTS

This project was undertaken in the Applications and Impact Division of the Canadian Climate Centre, Environment Canada. Mr. M.O. Berry provided project supervision and the report was edited by J. Masterton and B.F. Findlay.

REFERENCES

Gregory, S. 1963. Statistical methods and the geographer. Longmans, Green and Co. Ltd., London.

Lamb, H.H. 1970. Volcanic dust in the atmosphere; with a chronology and assessment of its meteorological significance. Philosophical Transactions of the Royal Society, London 266:425-533.

Mass, C., and S.H. Schneider. 1977. Statistical evidence on the influence of sunspots and volcanic dust on long-term temperature records. Journal of Atmospheric Science 34: 1995-2004.

Oliver, R.C. 1976. On the response of hemispheric mean temperature to stratospheric dust: an empirical approach. Journal of Applied Meteorology 15:933-950.

Pollack, J.B., O.B. Toon, C. Sagan, A. Summers, B. Baldwin, and W. Van Camp. 1976. Volcanic eruptions and climatic change: a theoretical assessment. Journal of Geophysical Research 81:1071-1083.

Schneider, S.H., and C. Mass. 1975. Volcanic dust, sunspots and temperature trends. Science 190:741-746.

Taylor, B.L. 1978. Volcanic eruptions and long-term temperature records: an empirical search for cause and effect. M.Sc. Thesis, University of Toronto. (A paper with the same title was published in Quarterly Journal of the Royal Meteorological Society 106: 175-199, 1980.)

Wexler, H. 1951. On the effects of volcanic dust on insolation and weather. Bulletin of the American Meteorological Society 32:48-51.

HISTORIC RECORDS

SEA ICE AND CLIMATIC CHANGE IN THE CANADIAN ARCTIC SINCE 1800

Moira Dunbar[1]

INTRODUCTION

Sea ice is potentially a useful indicator of climatic change. It is as it were visible sea temperature, and its presence or absence in a given area is a significant climatic factor. In practice, however, its value is greatly reduced by difficulty of exact measurement and by the scarcity of historical records. There are exceptions, notably the Icelandic records[2] of which Koch (1945) made such good use in his classic study of the East Greenland ice; there are also good records for the Baltic.

In the Canadian Arctic, there are few records until the nineteenth century, and for most areas records are so unevenly distributed both in time and place as to be of limited value. They are also difficult to relate to modern conditions and even to each other, owing to the wide variation in type of vessel used and the inevitable subjectivity of descriptions. However, the conclusion is inescapable that conditions in the nineteenth century were considerably more severe in at least some areas than they are now, and for one area, Baffin Bay, enough information is available to give concrete evidence of this. In this paper, I will discuss Baffin Bay and Parry Channel, adding a few remarks about other areas of interest.

Before 1800, voyages into arctic Canada were mostly restricted to the ice-free summer period; few expeditions wintered. And with one exception they concentrated on Hudson Bay and southeast Baffin Island. The exception was Baffin's voyage of 1616 - a remarkable achievement in which he circumnavigated Baffin Bay. Coasting up the west side of Greenland to 77°30'N, he discovered the open water at the head of the bay (later known as the North Water) and named Smith, Jones, and Lancaster sounds. But he made the serious mistake of

---

[1] R R #1, Dunrobin, Ontario K0A 1T0

[2] Ed. note: Ogilvie (1984) provides a current review of past climate and sea ice in this region.

reporting that there was no exit to the bay, and his discoveries were not followed up. It was not until two centuries later that serious exploration of the area, and of the archipelago lying to the west of it, began. At this time the British Navy, released from the Napoleonic wars and looking for ways to employ its ships and men, undertook the search for the Northwest Passage, which had been dropped by the merchant adventurers in the eighteenth century when Hudson Bay failed to yield an outlet to the west. In attempting this more northerly latitude, explorers of the Canadian Arctic really came to grips with sea ice for the first time.

Between 1818 and 1845 there were eight seaborne expeditions, five of them by the Baffin Bay route. The first, under John Ross, almost exactly duplicated Baffin's voyage - even to the statement that Lancaster Sound was an enclosed bay. However, Ross's second-in-command, Edward Parry, was not convinced of this and in the following year (1819), he proved it brilliantly by sailing all the way to Melville Island. Three more expeditions followed, the last and most famous being the Franklin expedition of 1845, the loss of which led to a period of intensive search, so that from 1848 to 1859, when the search ended, sea-ice information is available for nearly every year. The areas to be discussed in this paper, Baffin Bay and Parry Channel, were those traversed by the greatest number of vessels.

BAFFIN BAY SEA-ICE RECORDS

Data for Baffin Bay are more plentiful than for any other area because, not only did many explorers pass through it, but it also supported a large and thriving whaling industry throughout the nineteenth century. In the peak period (1830), as many as 91 ships were involved, of which 19 were lost. In 1817, just one year before Ross, a few whalers penetrated as far as Melville Bay for the first time. They saw, but did not enter, the North Water. In 1818 Ross found the open water to be full of whales, and from then on the whalers vied with each other to reach this rich area as early as possible. Although they left no extensive ice records, the dates of their voyage up the coast are in themselves a fair indication of the earliest possible date of passage for the type of vessel used. Given the pattern of break-up in Baffin Bay (Figure 1), the dates of coastal passage can be related in a reliable way to the progress of break-up. This subject is treated in more detail in Dunbar (1972).

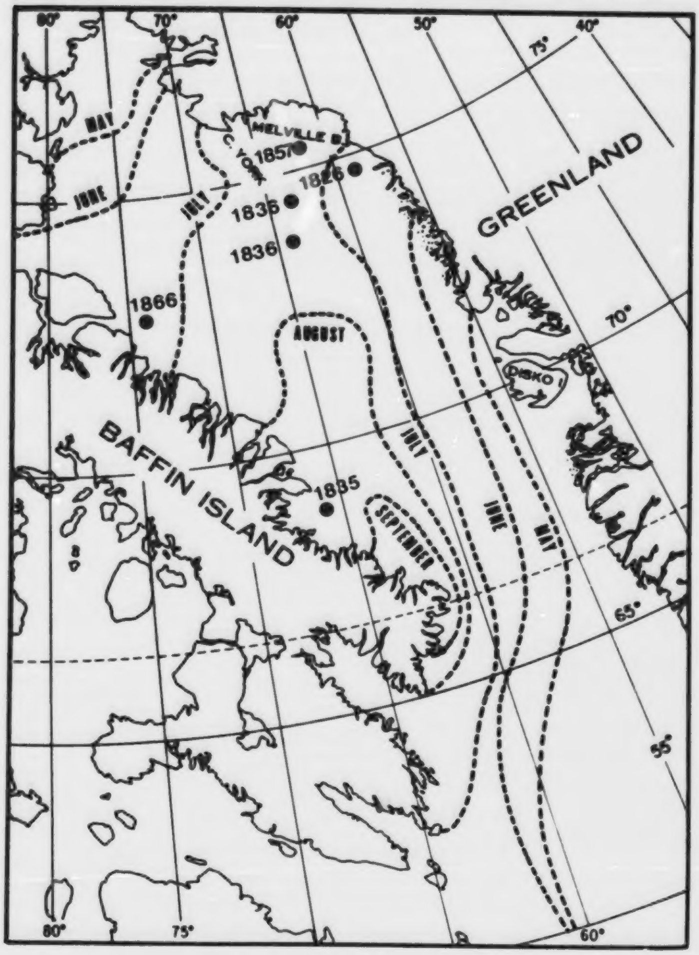

80°
70°
60°
50°
75°
40°
MAY
JUNE
JULY
MELVILLE B.
C. YORK
1857
1826
1836
1836
GREENLAND
1866
AUGUST
70°
DISKO I.
BAFFIN ISLAND
JULY
JUNE
MAY
1885
SEPTEMBER
65°
55°
80°
75°
60°

FIGURE 1: *Mean monthly limits of 4-5/10 ice cover for the years 1961-1970. The dated points show positions where ships were beset and forced to winter in ice (see text).*

Table 1 shows approximate dates of reaching the North Water by sailing ship for the periods 1817-59 and 1960-77. The first column is based on actual records, the cut-off date of 1859 being chosen because from then on steam vessels came into use, and it becomes more difficult to interpret the records. The second column attempts to relate air reconnaissance records of sea-ice concentration to sailing-ship capability. The criterion chosen was a concentration of less than five-tenths ice in the most difficult area. In general, this is believed to be less than the ice sometimes actually penetrated by ships, but on the other hand takes no account of other factors which delay sailing ships, like calms and adverse winds - so the figures are at best an approximation. This is probably the reason for the wider spread of dates in the first column compared to those in the second, which are based on ice concentration only. It is interesting that the period of opening for the greatest number of years is the same in both columns, late July, but this may be coincidence. Of much greater significance is the large number of years in which no ships reached the North Water at all in the nineteenth century. This really shows the difference between the first period and the second. No matter how you treat the figures, there is no way to arrive at a failure to reach the head of Baffin Bay in any year since aerial records have been kept (about 1950). Conditions in the 1950s were even more favourable than shown in Figure 1 for the 1960s (Dunbar 1972). The 1970s have proved to be similar to the 60s, but at most the amount of ice remaining throughout the summer is insufficient to hinder the progress of any ship, except occasionally off the southeast coast of Baffin Island.

If further evidence is required it can be found in the records of ships beset in the ice. In 1857, Leopold McClintock in the *Fox*, engaged in the Franklin search, was beset in Melville Bay in mid-August and was not released until late April of the following year, having drifted halfway across Baffin Bay, and thence down the middle to the ice edge in Davis Strait. This is the best known and best documented case of forced wintering in the Baffin Bay ice, but several whalers had similar experiences. In 1826, the whaler *Dundee* was beset off the coast of Greenland at about 74°15'N on 19 August, to be released like McClintock at the ice edge in April. On October 1, 1835, seven ships were beset in Home Bay, southeast Baffin Island. The following year, six were beset in northern Baffin Bay (73°14'N to 74°30'N), one of which, the *Thomas*, was lost. Another vessel was not released until 21 May. Finally, the steam whaler *Diana* was trapped off Pond Inlet in August 1866. Unable to escape, she was run into the pack to drift on 23 September off Scott Inlet. Another vessel with more power managed to escape to open water.

TABLE 1: *BAFFIN BAY - DATES OF REACHING NORTH WATER BY SAILING SHIPS.*

DATES	1817-1859 NUMBER OF YEARS	1960-1977 NUMBER OF YEARS[1]
June 21-30	1	-
July 1-10	3	5
July 11-20	2	3
July 21-31	8	7
August 1-10	4	2
August 11-20	2	1
August 21-31	1	-
Not at all	10	-
Insufficient data	12	-
Total	43	18

[1] This column attempts to relate air reconnaissance records of sea-ice concentration during 1960 - 1977 to sailing-ship capability.

The approximate positions of all these besetments (Figure 1) show that, although they are all within or close to the zone where the ice lingers longest, only one is in the area where it tends to remain all season today. Even here the trapping of seven vessels, though not impossible, is unlikely. In the other areas sailing vessels might well be beset in July, but they would be sure to be released in August or September.

Figure 2, taken from Dunbar (1972), shows dates of reaching various points in Baffin's progress up the Greenland coast in 1616, the whalers under sail, and for the 1950s and 1960s, the latter calculated in the same way as the data for 1960-77 in Table 1. It is inserted chiefly to show that Baffin, operating at the end of a relatively warm period, was quite a lot earlier than the whalers and not out of line with present conditions. I realize that actually Baffin's line should be moved back 11 days, because he would have been going by the Julian calendar. Also, of course, he may have struck a lucky year. The fact that both twentieth century lines slope at a more gradual angle than the other two suggests that I may have underestimated the capability of sailing ships. If this is so, the latter-day whalers in the 50s and 60s would have proceeded at a sharper pace up the coast and the contrast would have been more striking, both here and in Table 1.

PARRY CHANNEL SEA-ICE RECORDS

The evidence for Parry Channel is much scarcer. However, over a period of 37 years from 1818 to 1854, there were 14 in which ships operated in Parry Channel. However, for two of these, 1845 and 1846, the two years the Franklin expedition was in the area, we know only where the ships went, not when.

Parry Channel is divided into Lancaster Sound, Barrow Strait, Viscount Melville Sound, and McClure Strait, and as the normal approach was from the east it is not surprising that the first two have borne the most traffic. They are also the most open. Only two expeditions in the period covered penetrated beyond Barrow Strait, so data are insuffcient to have much meaning. Therefore, only Lancaster Sound and Barrow Strait are dealt with in detail. I consider the two bodies of water as being divided by a line running across the channel from the northeast corner of Somerset Island.

Table 2 gives the same data for these two areas as Table 1 does for Baffin Bay. Both seem to have experienced a shift to earlier opening in this century. The second column takes into account the necessity of navigating Baffin Bay in order to reach Lancaster Sound, so

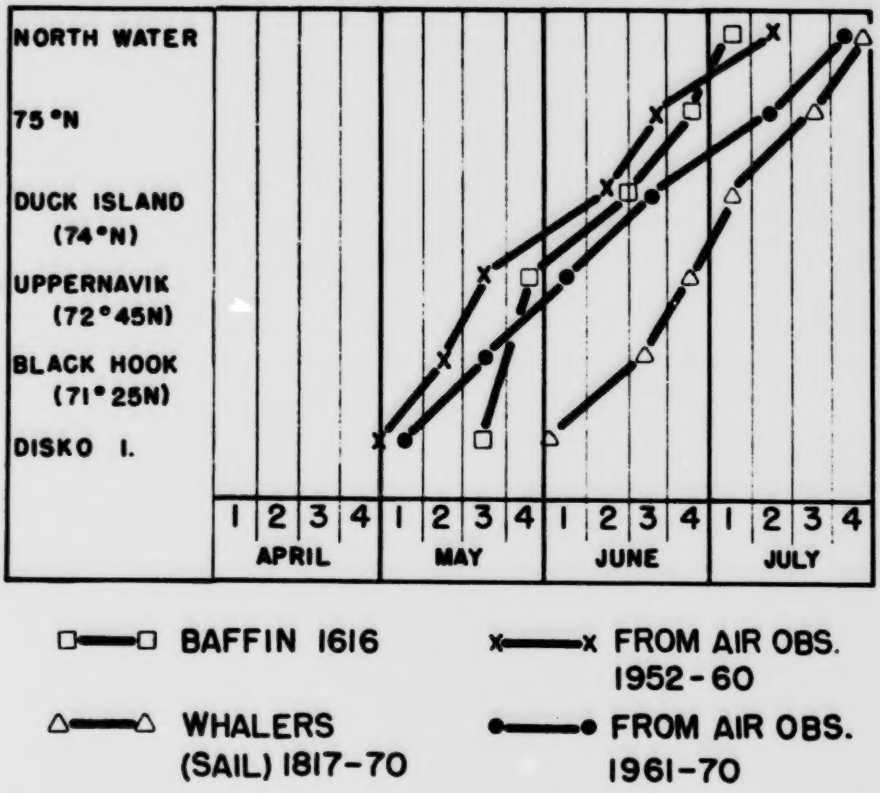


FIGURE 2: *Dates of first reaching various points on the west coast of Greenland.*

TABLE 2: *PARRY CHANNEL - DATES OF TRAVERSING BY SAILING SHIPS.*

DATES	LANCASTER SOUND		BARROW STRAIT	
	1818-1854 NUMBER OF YEARS	1960-1977 NUMBER OF YEARS[1]	1818-1854 NUMBER OF YEARS	1960-1977 NUMBER OF YEARS[1]
July 1-10	-	2	-	1
July 11-20	-	4	-	4
July 21-31	3	4	-	2
August 1-10	4	6	5	7
August 11-20	1	1	1	1
August 21-31	3	1	2	1
September 1-10	1	-	1	2
September 11-20	1	-	-	-
September 21-30	-	-	-	-
Not at all	-	-	1	-
No data	24	-	26	-
Total	37	18	(1846 traversed - date unknown)	

---

[1] See Table 1 for explanation of this column.

does not represent the actual date of opening of Lancaster Sound, which is often much earlier. The same was no doubt true in the nineteenth century.

Barrow Strait, both then and now, is a more difficult proposition than Lancaster Sound. As far west as Cornwallis Island, it is normally open as early as, or soon after, Lancaster Sound, but from a little to the west of Resolute Bay it breaks up much later and often only partially. Of the 10 years represented in the first column for Barrow Strait (Table 2) only five involve through-passages. There is also one other which does not appear in the table, because it cannot be dated - namely Franklin's passage in 1846 from his first wintering at Beechey Island to Peel Sound. His 1845 circumnavigation of Cornwallis Island, has been included on the basis of an educated guess based on the date of his last encounter with whalers in Baffin Bay. Thus we have a total of six through-passages: two (J.C. Ross in 1848, and Austin in 1850) intended passages that were stopped by ice at Griffith Island; two (Inglefield in 1853 and 1854) which never intended to go farther than Beechey Island; and one which failed to enter the strait at all (J.C. Ross, who was beset near Prince Leopold Island on 1 September 1849 and drifted with the pack through Lancaster Sound to Pond Inlet before being released).

Only two expeditions sailed west of Barrow Strait. The first was Parry's remarkable voyage in 1819, when he reached the entrance to McClure Strait in what was undoubtedly an above average open year. Released on 1 August 1820 from his winter quarters at Winter Harbour, he was unable to better his westing of the previous year and returned through Parry Channel to the east. The beginning of August is, of course, a very early date to attempt McClure Strait. Had he been in less haste to leave Winter Harbour it is conceiveable, though not very likely, that he might have completed the Northwest Passage. It would not be much more likely today.

The other expedition to reach Melville Island was Henry Kellett's in 1852. He wintered a little to the east of Parry at Dealy Island. Between these two voyages, two Franklin search expeditions (J.C. Ross and Austin) tried this route but were stopped by ice in Barrow Strait. In addition, McClure in the *Investigator* and Collinson in the *Enterprise*, starting from Bering Strait, sailed up Prince of Wales Strait in consecutive years and were turned back by ice at the entrance to Viscount Melville Sound. McClure then sailed south and west around Banks Island and entered McClure Strait from the west, reaching Mercy Bay on the north coast of Banks Island. Rescued by Kellett in his second year there, McClure and his crew completed the Northwest Passage on foot and by other ships, but the *Investigator* never left Mercy Bay.

None of these voyages, given the same ships, would be likely to be much more successful today. A calculation of the dates for sailing ships to reach southeast Melville Island for 1960-77 (as in Tables 1 and 2) gives one year in early August, three in late August, 11 in early September and two not at all. The early September years are by no means certain, so very likely, there would be more than two no-go years.

VICTORIA STRAIT AND GULF OF BOOTHIA SEA-ICE RECORDS

These two areas are of special interest, each because of the experience of one expedition. The first was the Franklin expedition, which was beset at the north end of Victoria Strait on 12 September 1846. The ships were abandoned in April 1848, having drifted only a little distance into the strait. Therefore, the ice did not break up in 1847, but it is not possible to say whether it did in 1848, as April is far too early for signs of break-up. The tradition arose that it was essential to use the narrow island-filled channels on the east and south sides of King William Island, which usually became sufficiently clear of ice. Whether this was really true at the time is hard to say, because, after Franklin, nobody attempted Victoria Strait until quite recently. It is certainly not one of the most navigable channels, but its record from 1960 to 1977 (Table 3) would have given Franklin a 60-65% chance to exit southward. However there were two two-year periods with no passage, so it is impossible to say whether there has been any improvement.

The same is true of the Gulf of Boothia. John Ross sailed down Prince Regent Inlet and wintered in a small bay in southern Boothia Peninsula in 1829. His expedition spent three winters in three different harbours within a distance of 15 miles (24 km) without being able to get out, and then abandoned their small vessel, the *Victory*, and retreated on foot to spend a fourth winter on the coast of Somerset Island before successfully escaping by boat to Lancaster Sound, where they were picked up by a whaler. For this area the 1960-77 figures are slightly less favourable than for Victoria Strait, giving Ross only a 56% chance of getting his ship out. Here too there are two two-year periods of continuous ice cover, and it is impossible to say whether conditions have improved or not.

TABLE 3: *DATE OF PASSAGE FOR SAILING SHIPS, 1960 - 1977.*

DATES	VICTORIA STRAIT (HEADING SOUTH)	GULF OF BOOTHIA (HEADING NORTH)
August 1-10	1	-
August 11-20	4	1
August 21-31	5	4
September 1-10	-	3
September 11-20	-	-
September 21-30	-	1
Doubtful	2	-
Not at all	6	-
No data	-	2

POLYNYAS

The polynyas, which exist today as areas of open water in winter or early spring, seem to have been very similar in the 1800s. The North water is well documented by the whalers and by several wintering expeditions in Smith Sound. The Franklin search was badly misled by the discovery of open water north of Cornwallis Island in early spring. This deflected the search from the southward route actually taken by Franklin, who had in fact followed his instructions to the letter. For a while the North water, together with polynyas found at the entrance to Kennedy Channel a little farther north, gave encouragement to a theory once popular that ice formed only close to the land, and that to the north of it was an open polar sea. This theory was not finally laid to rest until the 1870s, when the expeditions of Hall and Nares penetrated to the head of Nares Strait and beyond.

CONCLUSION

Based on good evidence for Baffin Bay and adequate evidence for Lancaster Sound and Barrow Strait, it can be definitely stated that ice conditions in these waters are considerably less severe at present than they were in the nineteenth century. For other areas in the Canadian Arctic the evidence is less clear, mainly because the nineteenth century sample is too small for comparison. Likely there has been less change in conditions in the more enclosed continental waters of the archipelago than in the large maritime area of Baffin Bay, situated close to the outer limit of ice cover. Both Baffin Bay and Lancaster Sound are further influenced by the early opening generated by the winter polynya of the North Water in Smith Sound. This, however, is known to have been a factor throughout recorded history and had as much influence in the 1800s as it does now.

This paper is based on a rather cursory study of the nineteenth century data. A more detailed study would undoubtedly reveal further interesting facts, but it is questionable whether the results would be worth the considerable effort involved, and unlikely that they would alter the general picture.

SUMMARY

Using available historical records of explorers and whalers, an attempt is made to show that, during the period of maximum exploration in the Canadian Arctic Archipelago (1818-1860 approximately), ice conditions were more severe than in the present century. Records are insufficient for positive proof, but for at least one area, Baffin Bay, there is enough information for a fairly clear picture to emerge. Evidence for the more continental waters west of Baffin Bay and Lancaster Sound is much less clear: perhaps less change has occurred there. The winter polynyas familiar today were also present in the nineteenth century.

REFERENCES AND SUPPLEMENTARY SOURCES

Belcher, E. 1855. The last of the arctic voyages;... L. Reeve, London. 2 Vols.

British Parliamentary Papers relevant to the arctic voyages. 1849-51 and 1854.

Canada, Atmospheric Environment Service. Ice reconnaissance data, 1960-77.

Collinson, Richard. 1889. Journal of HMS *Enterprise*,... Low, Marston, Searle and Rivington, London. 531 pp.

Cooke, Alan, and Clive Holland. 1978. The exploration of northern Canada. Arctic History Press, Toronto. 549 pp.

Davis, C.H. (editor) 1876. Narrative of the north polar expedition, USS *Polaris*. Washington. 696 pp.

Dunbar, Moira. 1972. Increasing severity of ice conditions in Baffin Bay and Davis Strait and its effect on the extreme limits of ice. In: Proceedings of the International Conference on Sea Ice, Reykjavik, Iceland, 1971. National Research Council of Iceland. pp. 87-93.

Kane, E.K. 1856. Arctic explorations:... Childs & Peterson, Philadelphia. 2 Vols.

Koch, Lauge. 1945. The East Greenland ice. Meddelser om Gr nland 130(3):1-374.

Lubbock, Alfred Basil. 1937. The arctic whalers. Brown, Ferguson, Glasgow. 483 pp.

McClintock, F.L. 1859. The voyage of the *Fox* in the arctic seas;...Murray, London. 375 pp.

McDougall, G.F. 1857. The eventful voyage of H.M. discovery ship *Resolute*... Longmans, London. 530 pp.

Nares, G.S. 1878. Narrative of a voyage to the Polar Sea... Low, Marston, Searle and Rivington, London. 2 Vols.

Ogilvie, A.E.J. 1984. The past climate and sea-ice record from Iceland, Part 1: data to A.D. 1780. Climatic Change 6 (2):131-152.

O'Reilly, B. 1818. Greenland, the adjacent seas, and the North West Passage... Baldwin, Craddock and Joy, London. 293 pp.

Osborn, Sherard. 1852. Stray leaves from an arctic journal... Longmans, Brown, Green and Longmans, London. 216 pp.

Parry, W.E. 1821. Journal of a voyage for the discovery of a Northwest Passage... in the years 1819-20,... Murray, London. 310 pp.

______. 1826. Journal of a third voyage... Murray, London. 186 pp. plus 151 pp. of Appendix.

Purchas, Samuel 1625. Hakluytus postumus, or Purchas his pilgrimes,... (for Baffin's journal).

Ross, John. 1819. A voyage of discovery... for the purpose of exploring Baffin's Bay... Murray, London. 252 pp.

______. 1835. Narrative of a second voyage in search of a Northwest Passage... Webster, London. 2 Vols.

EVIDENCE FROM HUDSON BAY REGION OF SEVERE COLD IN THE SUMMER OF 1816

A.J.W. Catchpole[1]

INTRODUCTION

A recent editorial by Schneider (1983) contributed to a debate which was initiated 25 years ago by Hoyt (1958) on the anomalous weather of 1816. It is, perhaps, inevitable that the questions first posed by Hoyt should continue to excite interest and yet remain unresolved. This debate touches upon large issues: 'the last great subsistence crisis of the western world' (Post 1977); the weather anomaly which has been characterized as 'the year without a summer' in New England and Western Europe (Stommel and Stommel 1979); a volcanic eruption which was virtually unparalleled in its intensity or effects in this millenium.

The separate elements of this debate have been submitted to detailed estimation and measurement:

(1) The great intensity of the dust veil emitted by the eruption of Mt. Tambora on the island of Sumbawa, Indonesia in 1815 was first apparent in its vivid optical effects described by eyewitnesses in Europe and North America (Post 1977). Recent research has yielded measures of the exceptional magnitude of the explosivity of the eruption (Newhall and Self 1982) and the high concentrations of acid deposited from the dust veil in Greenland (Hammer *et al.* 1980) and of particulate matter deposited in Antarctica (Thompson and Mosely-Thompson 1981).

(2) Measures of the degree of summer coldness in 1816 were revealed by regional reconstructions of annual mean summer temperatures undertaken by Manley (1974) for central England in the period since 1659, and by Landsberg *et al.* (1968) for eastern United States since 1738. Figure 1 illustrates the annual fluctuations of mean summer temperatures derived in each of these studies.

(3) In his detailed study of the economic depression and patterns of crop destruction in eastern North America and Western Europe in the summer of 1816, Post (1977) has provided a

---

[1] Department of Geography, University of Manitoba, Winnipeg, Manitoba R3T 2N2

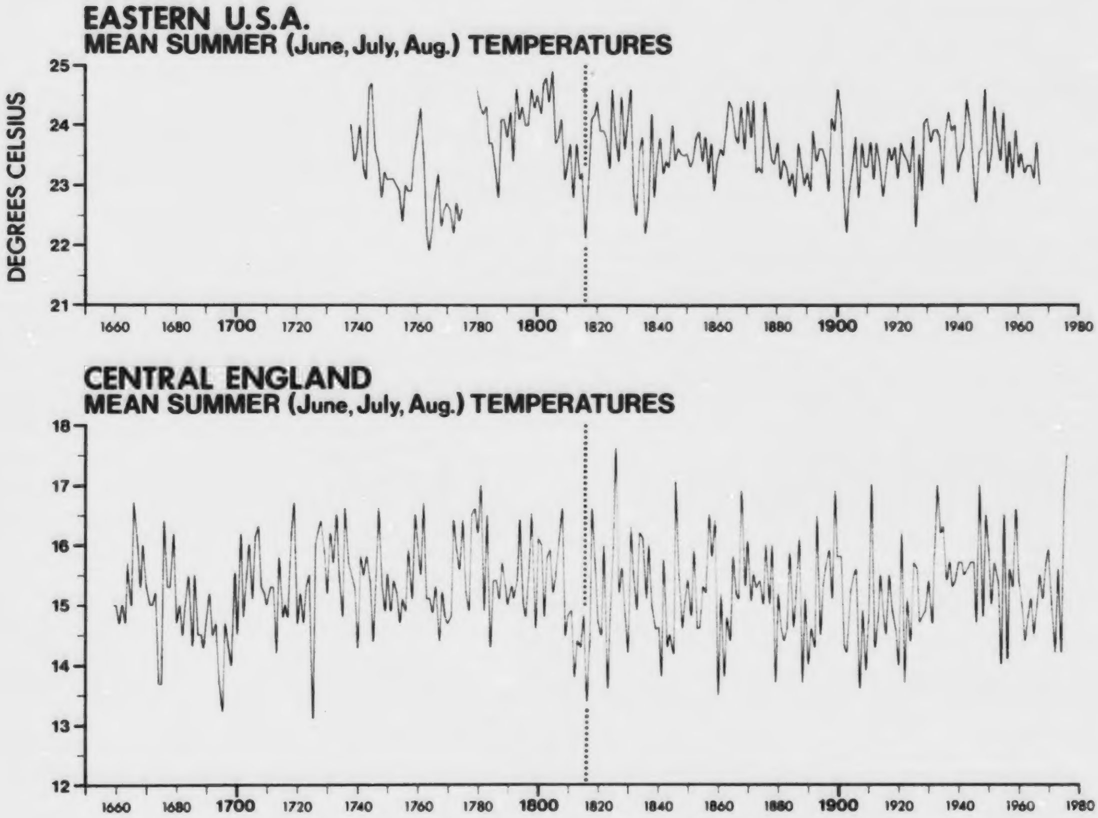


FIGURE 1: *Mean summer temperatures in eastern U.S.A. (Landsberg* et al. *1968) and in central England (Manley 1974).*

thorough reconstruction of the human aspects of events in 1816 and elaborated on Hoyt's (1958) earlier studies.

Thus the components of the problem have each been studied fully, but the fact remains that *these components have not yet been linked conclusively*. It is still not established whether the eruption of Mt. Tambora was the primary cause of the coldness of 1816 (Landsberg and Albert 1974) and it has not yet been shown conclusively that the crop losses of 1816 were indeed the result of the coincident coldness.

THE HUDSON BAY REGION IN THE SUMMER OF 1816

It is tempting to envisage the remote, subarctic Hudson Bay region as an empty wilderness in the summer of 1816, but this image belies the reality of its role as a hub of the fur trade. Scattered across the region were 19 Hudson's Bay Company fur trading posts (Figure 2) each housing a community of European traders ranging in population from eight to 38 persons. Two years previously, Company officers in London had directed that, where possible, temperature records were to be kept at its posts as a means of recording information which might relate to the food-producing potential of the local environment. By the summer of 1816, routine daily temperature measurements were being kept at nine posts: Churchill, Eastmain, Fort George, Great Whale River, Kenogamasee, Moose, Naosquiscaw, New Brunswick and York. Post journals containing records of daily occurrences at each of the posts shown on Figure 2 and of the temperatures observed at the nine listed posts have been preserved and are available for consultation in the Company's archives. Likewise, at sea, the Company's servants were active in the region. Hudson's Bay Company ships voyaged annually between London, England and the trading posts on the shores of the Bay. The log-books of these ships are also preserved today within the Company's archives. Five Company ships plied the waters of the Bay in 1816. Two of these, the *Eddystone* and the *Hadlow*, had entered the Bay in the summer of 1815 but had been prevented by sea ice from leaving in the fall of that year. These ships overwintered in Strutton Sound near Eastmain and had remained at anchor there until August 12, 1816. Meanwhile, the ship *Prince of Wales* and the brig *Emerald* entered the Bay on September 6, 1816 and reached Moose on September 20, 1816. On August 23, 1816 a coastal sloop left Great Whale River, bound for Fort George. This vessel encountered severe sea ice and, after a prolonged passage reached its destination on September 15, 1816.

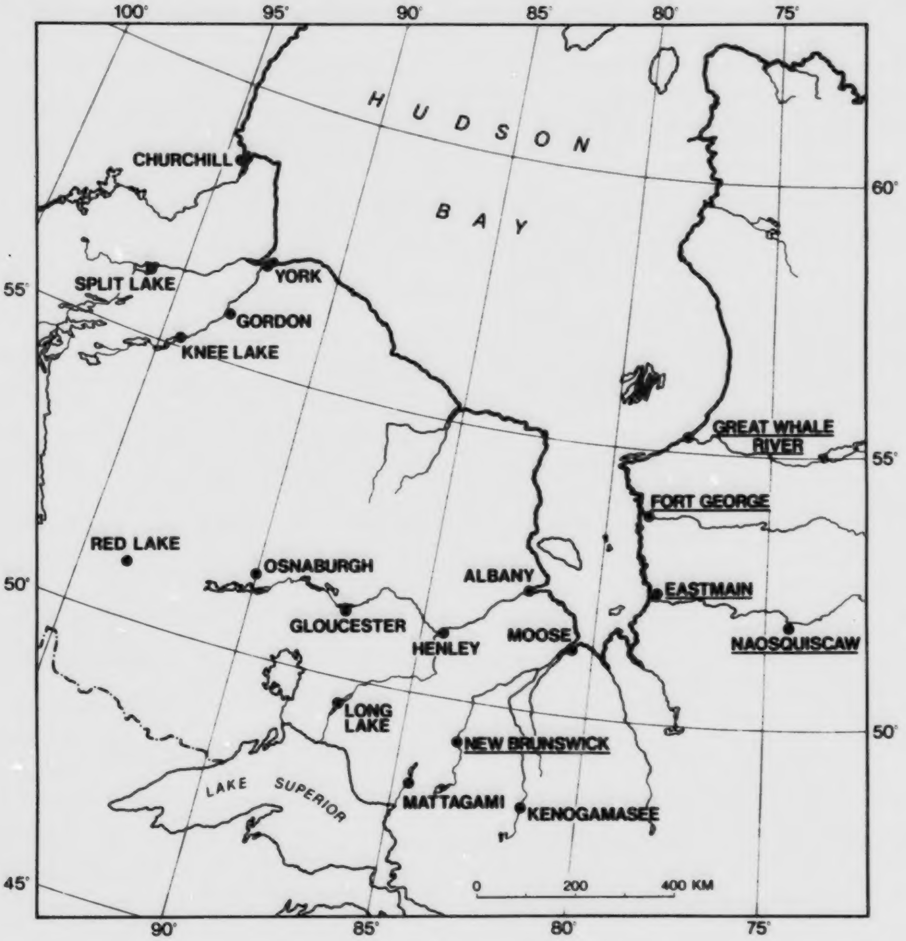

100°
95°
90°
85°
80°
75°
HUDSON
BAY
60°
55°
50°
45°
CHURCHILL
SPLIT LAKE
YORK
GORDON
KNEE LAKE
GREAT WHALE RIVER
FORT GEORGE
RED LAKE
OSNABURGH
ALBANY
EASTMAIN
GLOUCESTER
HENLEY
MOOSE
NAOSQUISCAW
LONG LAKE
NEW BRUNSWICK
LAKE SUPERIOR
MATTAGAMI
KENOGAMASEE
0
200
400 KM

FIGURE 2: *Locations of Hudson's Bay Company posts keeping journals in 1816. The underlined posts experienced severe garden-crop damage in 1816.*

In summary, the historical evidence of climatic and weather conditions in the Bay region is contained in 19 daily post journals, nine daily temperature records, four ships' log-books and a journal kept on one coastal sloop. A general discussion of the utility of these sources of historical climatic information is given in Catchpole and Moodie (1978) and in Catchpole (1980).

OBJECTIVES

It is instructive to recognize two general components in the weather anomalies which apparently occurred in the northern hemisphere in the summer of 1816. These comprise: (1) a prolonged, weak, general cooling, and (2) episodes of severe cold air outbreaks restricted to certain localities and possibly associated with the development of long waves in the westerlies and surface blocking action. The objectives of this paper are to use the available historical evidence in studies of both aspects of the weather anomaly in the Hudson Bay region. An important consideration is that this large, remote region yields evidence which is fragmented in terms of its physical type and sparse in its spatial distribution. It affords an insight into certain aspects of summer weather, but it cannot facilitate a comprehensive reconstruction of the prevailing weather conditions.

AMOUNT OF GENERAL COOLING

This account will consider general circumstances in Europe and eastern North America before focusing on the Hudson Bay region. Estimates of the degree of cooling may be derived from the records of air temperature which, by the second decade of the nineteenth century, were kept at many locations in Europe, the British Isles and eastern North America. These early observations were highly idiosyncratic with regard to: instrument design; exposure and site location; and observing procedure. This lack of standardization, together with the brevity or discontinuity of many records, militates against the direct use of raw temperature data in estimating the degree of cooling. For this purpose, I prefer to consider the reconstructed regional mean temperatures for central England (Manley 1974) and eastern U.S.A. (Landsberg *et al.* 1968). Figure 1 illustrates fluctuations in mean summer temperatures (June through August) in both of these regions. The summer of 1816 was approximately 2.0°C cooler than long-term normals in central England, and the equivalent cooling amounted to approximately 1.5°C in eastern U.S.A.

Similar estimates of regional mean summer temperatures are not, of course, available for the Hudson Bay region. However, the work of Parker *et al.* (1981) and Wilson (1982) furnish means whereby the depression of mean summer temperatures in 1816 at Great Whale River is estimated from two independent sources. Being independent in their origin, each of these estimates provides a check on the validity of the other.

Estimates Based on Tree-Ring Data

The estimate of summer cooling derived from the work of Parker *et al.* (1981) is based on tree-ring data collected from living white spruce (*Picea glauca*) at Cri Lake, a site located 4.2 km northeast of Great Whale River. The parameters measured were ring widths and maximum ring densities obtained by the application of x-ray densitometry. Parker *et al.* (1981) extracted from the raw data three components of temporal variation: (1) the growth trend occurring over the whole time series from pith to bark ("A" component); (2) short-term variations of period length between 10 years and the age of the tree ("B" component); and (3) year-to-year fluctuations ("C" component). A report from Parker *et al.* (1982) indicated the existence of a high correlation between mean summer temperatures at Great Whale River and year-to-year fluctuations in maximum ring density at Cri Lake. This prompted me to undertake an analysis of the relationship between mean summer temperatures at Great Whale River and the Cri Lake tree-ring data. In this context, summer is defined as the period May through October, and mean summer temperatures were calculated for 41 years from 1926 through 1975. An initial correlation analysis was applied with summer temperature treated as the independent variable and both the "B" and "C" components of ring-width and maximum ring density treated as the dependent variables. I found that the only significant correlation existed between mean summer temperature and the "C" component of maximum ring density ($r = +0.71$; significance level = 0.00001).

A step-wise multiple regression analysis was then applied in order to derive an estimate of mean summer temperature from the measured maximum ring density. This yielded the estimation equation:

$$[1] \qquad \overline{T}i = \frac{45.488}{3.162 - Di - 0.257Di^{-1}} - 17.8$$

where:

$\overline{T}i$ = mean summer temperature °C in Year i,
Di = maximum ring-density "C" component in Year i,
Di^{-1}= maximum ring-density "C" component of the year preceeding Year i.

Table 1 lists the estimated mean summer temperatures for Great Whale River during the period 1701 to 1925. The mean value for the entire period of record was 6.1°C with a standard deviation of 1.3°C. The mean summer temperature of 1816 (4.4°C) was thus 1.34 standard deviations below the mean. It is noteworthy that the degree of summer cold in 1817 was apparently considerably more severe than that of 1816, since the mean of that year 3.0°C was 2.4 standard deviations below the long-term mean.

Estimates Derived from Temperature Measurements

In a seminal study of the quality of the temperature data measured by Hudson's Bay Company personnel at Great Whale River, Fort George and Eastmain from 1814 through 1821, Wilson (1982) has laid the foundation for the scientific interpretation of these raw measurements. This study has yielded estimates of mean monthly temperatures during the summer (May through October). These estimates are corrected for instrument, exposure, site and procedural irregularities in the raw data. In her conclusion Wilson (1982, p. 204) made the following observation:

> "There is evidence that the temperature observations for Whale River, Big River [Fort George] and Eastmain from 1814 to 1821 were taken with care, and are reliable and consistent within the limitations of the instrumentation, sites, instrument exposure and observing practices of the period. It is believed that in their corrected form they are accurate to within acceptable limits of error compared with the modern series at Great Whale, Fort George and Eastmain. As such, they indicate that the period was generally cooler than recent times, and that the summers of 1816 and 1817 were colder than any on modern record."

The Great Whale River summer temperature record extends, with interruptions, from May 1815 to August 1816. Subsequently records in this general vicinity were kept at Fort George, approximately 190 km distant to the southwest, from October 1816 through June 1820. Using these corrected monthly mean temperatures, I calculated mean summer temperatures (May through October) for the Great Whale River/Fort George region in 1816, 1817 and 1819. This

TABLE 1: *ESTIMATES OF MEAN SUMMER (MAY THROUGH OCTOBER) TEMPERATURE (^{o}C) AT GREAT WHALE RIVER, 1701-1925, DERIVED FROM MEASUREMENTS OF TREE RING MAXIMUM DENSITY ("C" COMPONENT) AT CRI LAKE,[1] QUEBEC.*

YEAR	0	1	2	3	4	5	6	7	8	9
1700	-	7.3	6.0	6.8	6.0	5.8	6.5	8.0	5.2	3.5
1710	7.1	6.6	7.1	6.3	5.1	8.6	3.8	6.6	5.4	5.8
1720	5.0	5.6	7.3	7.1	6.3	6.1	5.0	5.6	6.5	8.6
1730	4.9	4.7	5.9	7.5	5.8	6.5	6.8	5.7	7.2	6.8
1740	4.1	5.7	5.9	6.7	6.9	6.3	4.6	6.1	7.3	7.2
1750	4.4	7.9	6.7	4.8	7.8	4.8	7.8	4.3	5.1	5.8
1760	5.7	6.1	5.4	8.8	8.1	3.7	6.8	5.0	6.4	7.6
1770	7.6	4.9	5.0	4.8	6.8	5.4	7.1	5.7	6.7	6.9
1780	4.8	7.2	8.3	7.0	4.0	6.1	5.9	4.4	7.0	8.4
1790	5.2	6.2	4.7	6.8	5.9	7.3	5.1	6.8	7.4	7.3
1800	3.6	5.9	5.8	5.5	8.0	8.3	5.5	4.3	5.5	5.6
1810	8.7	6.6	5.9	7.3	4.8	6.3	4.4	3.0	8.9	5.0
1820	7.5	8.0	5.4	5.8	4.6	6.1	8.5	4.2	7.1	6.7
1830	5.5	7.4	5.2	5.3	7.5	5.2	5.2	7.2	4.5	6.3
1840	8.5	7.7	4.5	4.1	7.2	4.4	7.6	6.0	5.8	6.8
1850	6.9	6.7	7.1	3.7	5.3	5.1	8.4	6.1	8.1	4.8
1860	6.9	6.4	4.5	5.0	6.8	5.0	7.2	7.5	5.9	4.5
1870	7.9	3.7	6.6	6.9	5.0	6.7	6.9	6.5	6.3	6.5
1880	4.6	7.6	7.1	5.7	3.7	4.4	8.3	7.8	4.1	7.4
1890	5.9	4.0	6.9	8.0	4.6	6.8	6.7	6.3	5.3	6.7
1900	6.0	7.4	5.7	6.3	5.6	4.7	6.2	6.5	6.5	5.3
1910	5.0	10.1	3.8	5.5	5.2	6.2	8.4	5.9	3.2	8.2
1920	7.4	6.1	7.4	4.0	8.0	4.6	-	-	-	-

---

[1] Tree-ring data from Parker *et al.* (1981).

calculation involved the substitution of one missing value (September 1816) using the mean September temperature at Great Whale River in the modern period, 1926-1975. Figure 3 presents the findings and compares mean summer temperatures based on temperature measurements with those derived from the tree-ring data. The vertical and horizontal axes are both calibrated in mean summer temperatures. The information plotted against the vertical axis includes: (1) the long-term mean and standard deviation of mean summer temperatures observed at Great Whale River in the period 1926-1975; and (2) the mean summer temperatures derived from Wilson (1982) for 1816, 1817 and 1819 at Great Whale River/Fort George. The information plotted against the horizontal axis includes: (1) the long-term mean and standard deviation of mean summer temperatures estimated from the Cri Lake maximum density record in the period 1701-1925; and (2) the mean summer temperatures estimated thus at Great Whale River in 1816, 1817 and 1819.

The 1816 mean based on temperature measurements was 3.2°C, an amount which is 2.5 standard deviations below the mean summer temperature (6.2°C) observed at Great Whale River in 1926-1975. The following year is again seen to be colder than 1816 since the measured mean summer temperature (2.7°C) in 1817 was 2.94 standard deviations below the 1926-1975 mean. It is apparent (Figure 3) that the vicinity of Great Whale River experienced general cooling in the summer of 1816 comparable to that indentified in central England and eastern United States, and that even more pronounced general cooling occurred at Great Whale River in 1817.

COLD AIR OUTBREAKS

A second general component identified in the weather anomalies in the northern hemisphere in the summer of 1816 was the occurrence of cold air outbreaks which brought severe cold to certain localities. These outbreaks were possibly associated with the development of long waves in the westerlies and surface blocking action. This component of the weather anomalies is revealed by three categories of evidence:

(1) Daily temperature observations which demonstrate that the summer of 1816 was punctuated by episodes of exceptional cold. Thus, a series of cold waves in New England was described as follows by Stommel and Stommel (1979, p. 176).

> "The first of three unseasonably cold waves moved eastward into New England early on June 6. The cold and wind lasted until June 11, leaving from three to six inches of snow on the ground in northern New England.

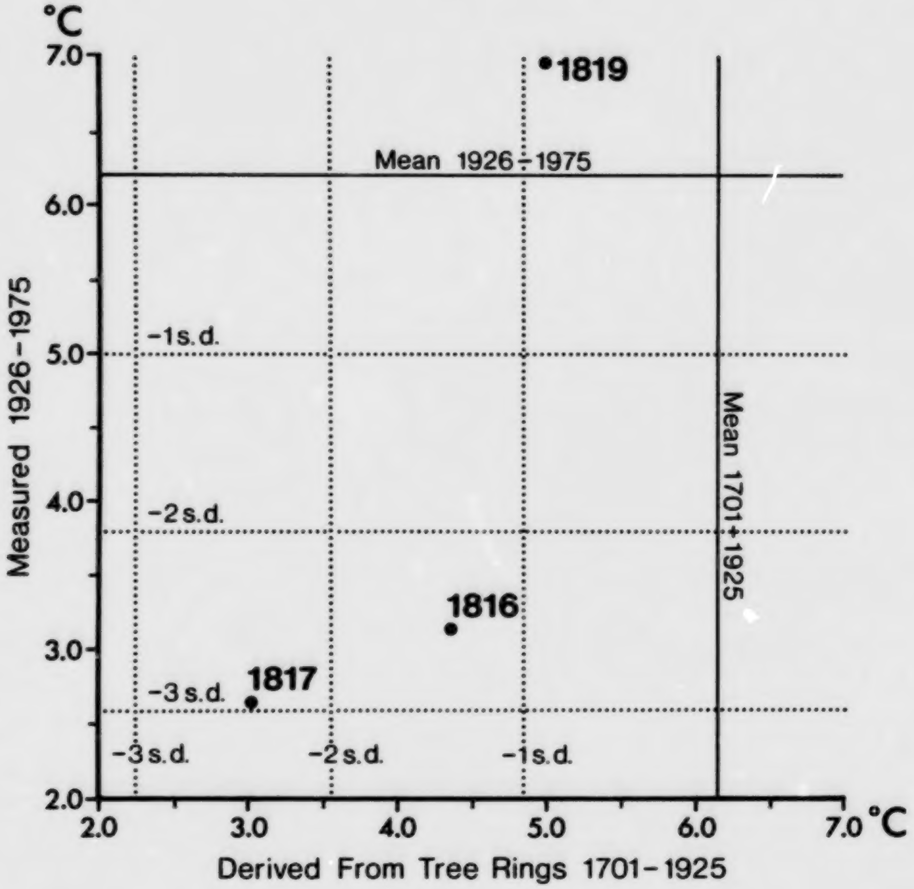


FIGURE 3: *Mean summer (May through October) temperatures at Great Whale River/Fort George in 1816. The values plotted against the vertical axis are derived from temperature measurements made at the Hudson's Bay Company posts (corrected values from Wilson 1982) and are compared with the long-term mean and standard deviation observed at Great Whale River in 1926-75. The values plotted against the horizontal axis are estimated from tree ring maximum densities measured at Cri Lake (Parker* et al. *1981). These values are compared with estimates of the mean and standard deviation derived from tree rings in the period 1701-1925.*

> A second killing frost struck the same areas on July 9 and a third and fourth on August 21 and 30, just as the harvest of twice-ravaged crops was about to begin. The repeated summer frosts destroyed all but the hardiest grains and vegetables.";

(2) The spatial patterns of crop destruction which reveal that some parts of Europe and North America experienced severe losses while others were relatively unscathed (Post 1977). Severe losses occurred in New England and southward along the Atlantic seaboard, but regions to the west of the Appalachians were much less affected. Likewise, Western Europe suffered severely while Scandinavia and the Russian plain avoided the worst effects;

(3) The reconstructed mean surface pressure pattern of July 1816 (Lamb and Johnson 1966) for regions bordering the North Atlantic indicates a northerly airflow over eastern North America and blocking action over Western Europe (Figure 4).

The Hudson Bay region provides three categories of evidence which are indicative either of cold air outbreaks, or of the prevalence of northerly airflow with which these outbreaks may have been associated. These include daily temperature observations and spatial patterns of garden-crop destruction which are analogous to two of the general types of evidence available in Europe and eastern North America. The third category of evidence in the Bay region involves patterns of sea-ice distribution and movement.

Daily Temperature Observations

In the summer of 1816, daily temperature measurements were made at nine of the posts shown in Figure 2: Churchill, Eastmain, Fort George, Great Whale River, Kenogamasee, Moose, Naosquiscaw, New Brunswick and York. The longest records were kept at Moose (May 1 to September 30) and at Great Whale River (May 1 to August 20). Usually the temperature records are limited to May and June. The information available in this study comprises the raw daily observations, and no effort has been made to standardize the data for differing times of observation, instrument designs, exposures, sites of observation or any of the other factors which degrade the quality of temperature readings. Figure 5B illustrates the daily fluctuations exhibited by the raw data observed at Kenogamasee and New Brunswick. For comparative purposes, these data are juxtaposed with average daily temperatures observed at New Haven, Connecticut and Williamstown, Massachusetts (Figure 5A). The vertical bar delimits the 'first of the three unseasonably cold waves' identified by Stommel and Stommel (1979) in New England. There are clear indications, even from the raw daily temperature data, that this cold wave which brought June snowfall to New England also swept over the Bay

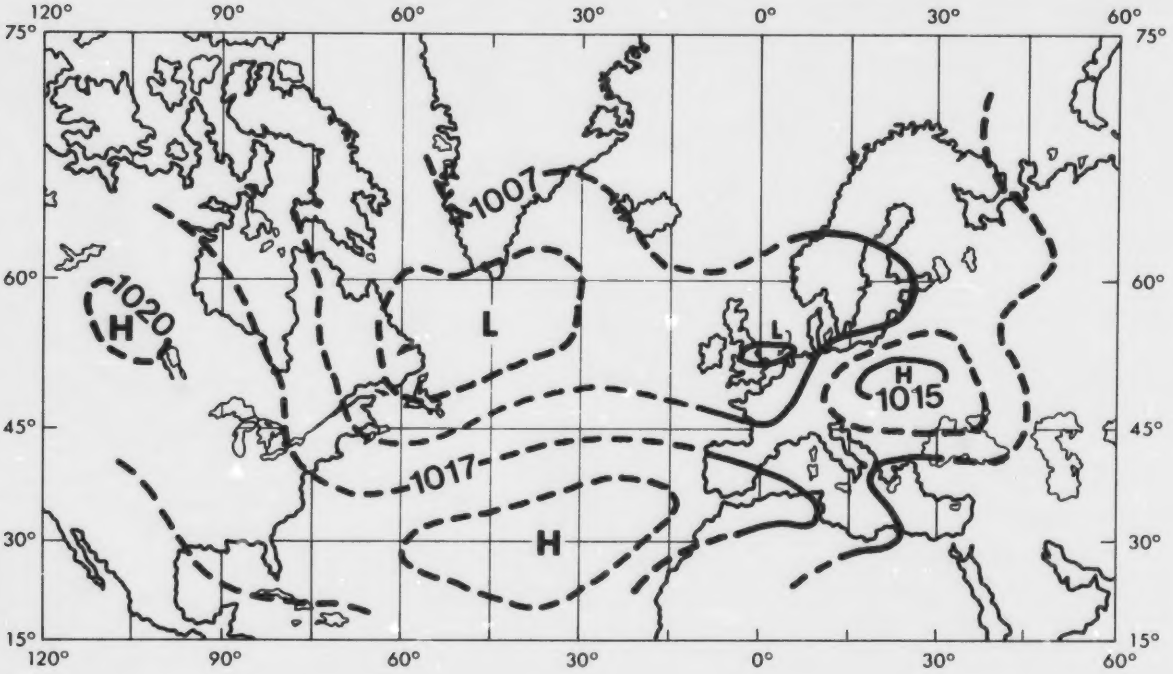

120°
90°
60°
30°
0°
30°
60°
75°
60°
45°
30°
15°
1007
1020
H
L
L
H
1015
1017
H

FIGURE 4: *Mean sea-level isobars (mb) in the vicinity of the North Atlantic Ocean, July 1816 (Lamb and Johnson 1966).*

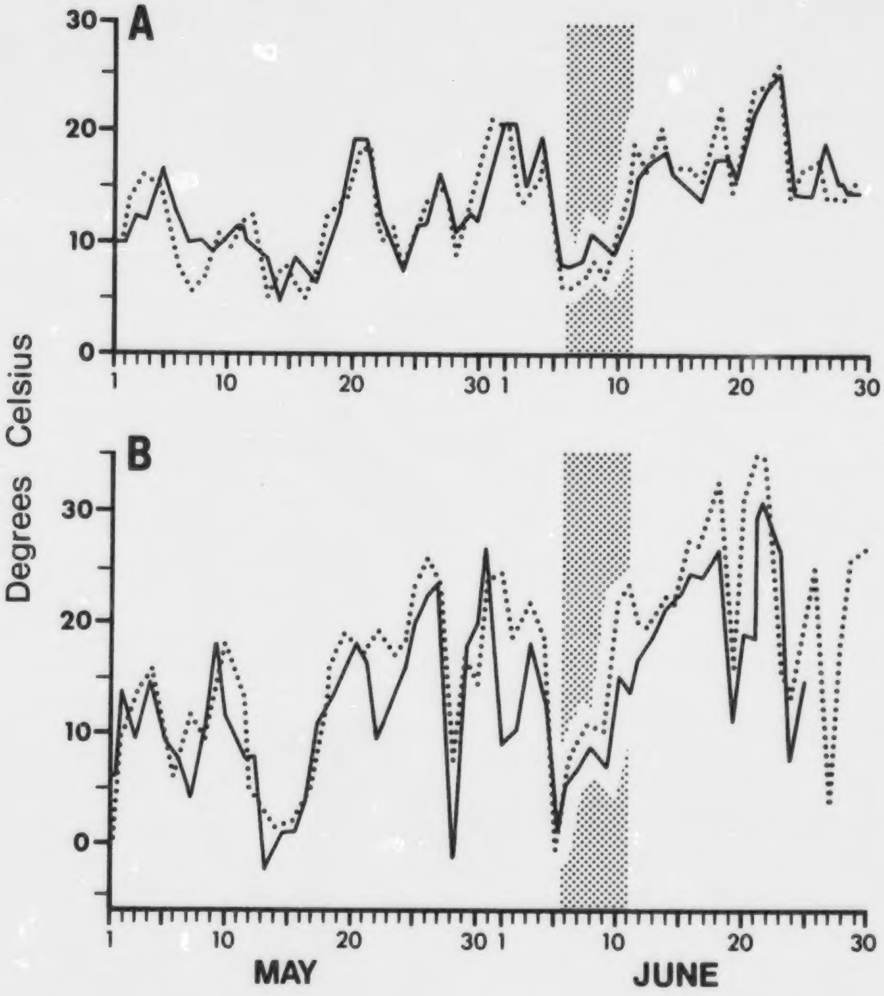


FIGURE 5: *(A) Average daily temperatures at New Haven, Connecticut (solid) and Williamstown, Massachusetts (dashed) in May and June 1816; (B) temperatures observed at 12:00 noon local time at Kenogamasee (solid) and New Brunswick (dashed) in May and June 1816. [Data in Figure 5A are from Stommel and Stommel (1979); those in Figure 5B are from Hudson's Bay Company, Kenogamasee journal, B99/a/18, fol. 6, 1816; and New Brunswick journal, B145/a/32, fol. 6-7, 1816].*

region. A noteworthy feature of Figure 5B is the relative homogeneity displayed by the daily temperature fluctuations at Kenogamasee and New Brunswick. This indicates that measurement errors in these unstandardized temperature readings are small in relation to the large amplitude fluctuations encountered on a day to day basis. Presumably, therefore, *there is scope for synoptic analyses*[1] of the daily temperatures observed in this network of nine trading posts in the summer of 1816.

Garden-Crop Destruction

A general feature of the way of life at the Hudson's Bay Company posts was the practice of gardening. In 1816, this was not a whimsical diversion from the routine work of the post, but an activity which had been vigorously pursued for over a century as an element of the Company's official policy (Moodie 1978). Shortly after the inception of its activities in 1670, the Company deliberately encouraged gardening at its Bayside posts with the objectives of raising locally provisions which were expensive to import from England and of fostering the health of the overseas servants. Of course, in these harsh subarctic environments the produce of the gardens contributed an almost negligible portion of the total foodstuffs consumed on the Bay. Nevertheless, what the products of gardening lacked in volume, they made up for in kind as a remedy to scurvy:

> "Despite many attempts, the gardens on Hudson Bay did not have any impact on the costs of provisioning the Company's settlements. They presisted in this inimical environment because of the contribution which they made to the diet, a contribution which, prior to the use of orange and lime juice as antiscorbutics, was virtually vital to the health of the traders. Despite the vagaries of the harsh climate, moreover, the gardens could in large degree be depended upon in this important role. The frequent failure of various vegetables to mature in the cool summers of the littoral did not render them useless as antiscorbutics, for they could still be eaten as greens."
>
> Moodie (1978 p. 59)

The Hudson's Bay Company records contain daily journals kept in the summer of 1816 at each of the 19 posts shown in Figure 2. In addition, journals are available from four posts located to the west of the Hudson Bay region. All of the daily accounts written in each journal between May 1 and October 31, 1816 were scrutinized with the objective of identifying descriptions of garden-crop destruction attributable to adverse weather conditions. A distinction was then drawn between delays in crop germination due to late frosts or snowfalls in early summer and evidence of total crop losses due to their failure to develop beyond initial germination or come to maturity. These losses were caused by the recurrence of episodes of severe cold in mid-and late summer, and it is noteworthy that they were restricted to the southeast quadrant of the Bay region. Thus crop losses were reported at Great Whale River, Fort George, Eastmain, Naosquiscaw and New Brunswick:

---

[1] Ed. note: See Wilson's paper "Daily Weather Maps for Canada, Summers 1816 to 1818 - a Pilot Study" in this volume.

> "The gardens at [Great] Whale River have not produced a single root of vegetables of any kind whatever."
> Hudson's Bay Company, Fort George journal, B77/a/3, fol. 5, October 10, 1816.
>
> "None of the grass or anything else has come to perfection this season. Continual frost and snow throughout the summer, has been a great impediment to all kinds of vegetation. Not so much as a berry of any kind is scarcely to be seen, which on more favourable seasons are found here, to grow spontaneously in great abundance."
> Hudson's Bay Company, Fort George journal, B77/a/3, fol. 3, September 27, 1816.
>
> "I have got no vegetables whatever, the plants that were sent so opportunely from Moose last summer are no larger than when planted, as for Potatoes, there is not the smallest branches to see above the ground."
> Hudson's Bay Company, Eastmain journal, B59/a/96, fol. 10, October 13, 1816.
>
> "The repeated frosts has destroyed the potatoes. The leaves are all gone, and nothing but the naked stalks remain."
> Hudson's Bay Company, Naosquiscaw journal, B.143/a/15, fol. 10, September 4, 1816.
>
> "I set to mowing down the oats, barley in the Park, as the season is pretty far spent, but the grain has come to nothing - the straw however will come in for the cattle."
> Hudson's Bay Company, New Brunswick journal, B145/a/34, September 20, 1816.

The spatial pattern of garden-crop destruction may indicate that while the eastern part of the Bay region received the full force of cold air outbreaks, areas further west escaped their worst affects. This is quite consistent with the Lamb and Johnson (1966) reconstruction of the mean July 1816 surface pressure pattern which shows a pronounced northerly airflow over the eastern Bay region while the western region is dominated by a high pressure cell over the central prairies (Figure 4).

Summer Sea-Ice Distribution and Movement

The prime basis upon which inferences are made concerning the atmospheric circulation is the pattern of late summer sea-ice distribution in Hudson Strait, Hudson Bay and James Bay. This information is mainly derived from the log-book of a ship sailing into Hudson Bay from England, but subsidiary sources include the log-book of a ship which overwintered in James Bay in 1815-1816, a journal kept on a coastal sloop sailing in James Bay and the post journals kept at Great Whale River and Eastmain.

Sea-Ice Dispersal and Atmospheric Circulation

The summer dispersal of sea ice in subarctic and arctic waters is a response to a multitude of factors including radiative heat exchange, sea currents, and the ebb and flow of tides. A major control of ice dispersal is exercised by surface winds through their advective and frictional affects, and it is through this relationship that ice provides evidence of wind conditions. Despite substantial recent research into summer sea-ice dispersal in eastern arctic waters (Crane 1978; Danielson 1971; Dey 1980; Jacobs and Newell 1979; Keen 1978; Markham 1981; Maxwell 1982; and Walsh and Johnson 1979), knowledge of the process and its controls is still highly uncertain. This stems from the geographical complexity of northern waters, the brevity of the period of aerial and satellite reconnaissance of sea ice and the sparsity of arctic meteorological observations. It is against this background of uncertainty that I use sea ice as evidence of atmospheric circulation in the summer of 1816. The inferences drawn in this study are speculative and their validity cannot be demonstrated with the information now available.

Several present-day studies have linked rates of summer sea-ice dispersal in Baffin Bay, Davis Strait and the Labrador Sea to atmospheric circulation patterns. One finding is that the rate of dispersal is related to the longitudinal position of the axis of the trough in the upper westerlies, which normally extends from eastern North America to eastern Asia (Barry 1981). A westward displacement of this axis encourages incursions of southwesterly surface winds in the eastern Hudson Bay region. This permits mild air to be advected northwards, it retards the southward drift of ice from Baffin Bay and Davis Strait, and from Foxe Basin into Hudson Strait. Thus, a relatively early ice dispersal occurs in Hudson Strait. By contrast, an eastward displacement of the axis of the upper trough allows cold northwesterly winds to prevail over the region. These winds delay melting while increasing the southward flow of ice into Hudson Strait, thus retarding ice dispersal in the Strait. These relationships were identified by Keen (1978) in Baffin Bay and by Crane (1978) in the Davis Strait-Labrador Sea region.

Summer sea-ice dispersal in the western part of Hudson Bay normally spreads southward from Southampton Island, while in the east it spreads northward from James Bay (Danielson 1971). This pattern of dispersal is determined by the counterclockwise sea currents in the Bay, by the prevalence of northwesterly winds circulating cyclonically into the Icelandic low and by the northward progression of radiative heating. Consequently, the last ice in Hudson Bay is usually found in the southwest quadrant to the west of Cape Henrietta Maria.

Sea Ice in 1816

A previous study has derived annual indices of summer sea-ice severity for Hudson Strait in the period 1751 to 1870. The log-books of the Hudson's Bay Company ships, which comprise the source of this information, were described in Catchpole (1980) and the method of analysis of the log-books was outlined in Catchpole and Ball (1981) and described fully in Catchpole and Faurer (1983). The index of ice severity was based on the durations of the annual westward passages of the ships through Hudson Strait and the frequency with which the ships undertook specific navigational manoeuvres when they encountered ice. From this study, 1816 emerged as the year with the highest summer sea-ice severity index in the whole period of record.

It is, of course, possible that this result was merely reflective of the coincident general cooling in the northern hemisphere, but an analysis of ice conditions, based upon the information in the log-book of the brig *Emerald* (Hudson's Bay Company, log-book of Brig *Emerald*, C1/324, 1816) revealed features of ice distribution and movement which are probably indicative of prevailing northerly or northwesterly winds (Catchpole and Faurer 1983). This analysis was based on the route followed by the *Emerald* through the Strait and the ice conditions encountered at various stages along the route. Figure 6 reconstructs this passage indicating the ship's progress at weekly intervals between its first sighting of Resolution Island and its arrival at Moose Factory. Also shown on Figure 6 is the route followed by the *Emerald* between October 5 and October 13. This effort to return to

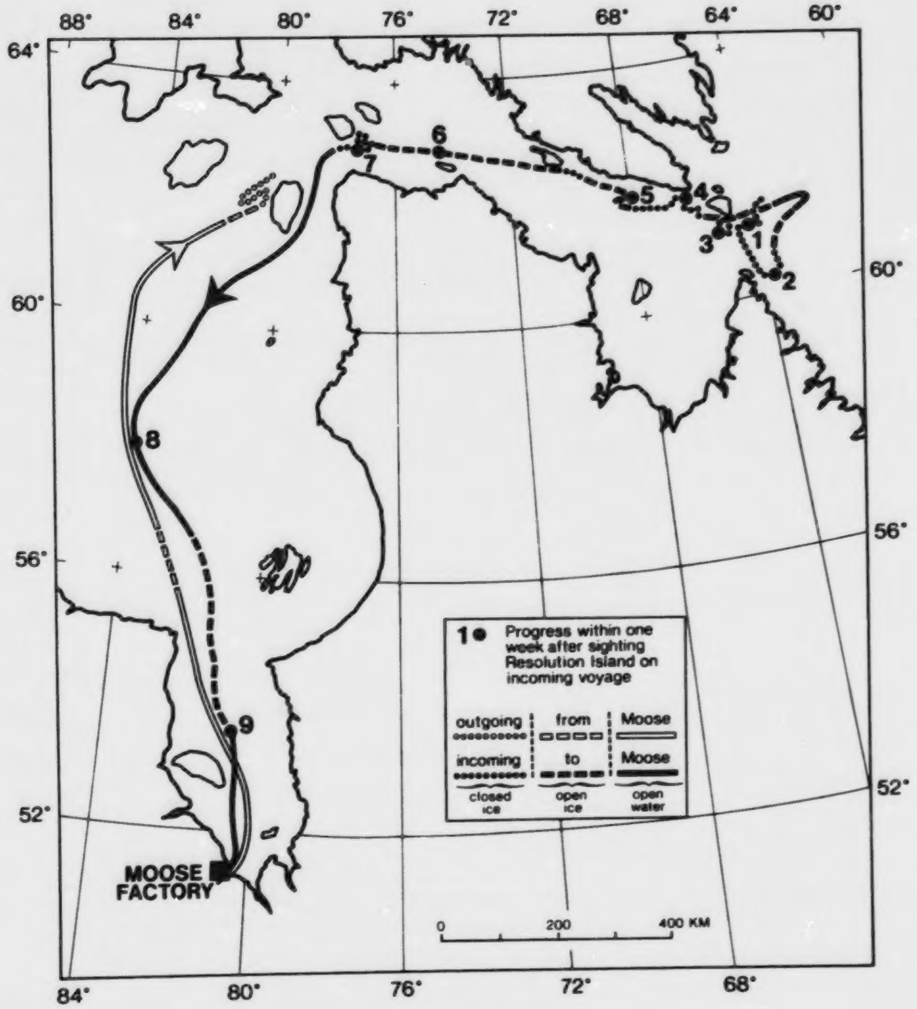

88°
84°
80°
76°
72°
68°
64°
60°
64°
60°
56°
52°
60°
56°
52°
1
2
3
4
5
6
7
8
9
MOOSE FACTORY
1● Progress within one week after sighting Resolution Island on incoming voyage
outgoing
from
Moose
incoming
to
Moose
closed ice
open ice
open water
0
200
400 KM
84°
80°
76°
72°
68°

FIGURE 6: *Passage of the brig* Emerald *to and from Moose Factory in 1816. The incoming passage is plotted from July 17, when Resolution Island was first sighted, to September 20 when the ship reached Moose Factory. The outgoing passage commenced on October 5 and terminated on October 14 when impassable ice was encountered. The return passage to Moose Factory is not shown. The numbers indicate progress at weekly intervals during the incoming passage.*

England was aborted when impassable ice was encountered at Mansel Island near the entrance to Hudson Strait. At this point the *Emerald* returned to James Bay and then overwintered in the Bay. General sea-surface conditions are identified on Figure 6. The designation of ice conditions into the categories of open and closed is a subjective classification intended to simplify the miscellany of descriptive commentaries appearing in the logs. It is noteworthy that when the ship was in closed-ice conditions its motion often appeared to be dictated by normal patterns of ice drift and deviated from a direct route through the Strait. This implies that, in closed ice, the ship was compelled to grapple with ice or was beset in ice. Likewise, in open ice, the ship was generally able to progress along a direct route.

A detailed explanation of the procedure used to reconstruct the ship's route through Hudson Strait, shown in Figure 6, and a detailed interpretation of the implications of this passage regarding sea-ice and atmospheric circulation is given in Catchpole and Faurer (1983). The *Emerald* first approached Resolution Island on July 17, but became caught in a rapid southward flow of ice and, in the first two weeks thereafter drifted south past Button Island off the north coast of Labrador. This was apparently the only occasion in the period of record in which a Company ship drifted south in this manner across the eastern entrance to the Strait. It is reasonable to assume that during this drifting the *Emerald* was beset in a vigorous flow of ice southward from Davis Strait in the Canada Current. After 14 days the *Emerald* escaped from the closed ice, sailed eastwards into open water and again approached Resolution Island on August 3. On this occasion the ship successfully entered the Strait, but, after rounding Resolution Island, became enclosed in ice described in the log-book as moving rapidly westward. This was probably indicative of a vigorous flow of ice in an eddy of the Canada Current which generally moves westward along the Baffin Island coast from Resolution Island to Big Island in the middle of the Strait. A significant feature is that open water was observed to the north when the ship was drifting westward in closed ice during the fourth and fifth weeks and later when the ship was approaching Big Island. Under these circumstances, open water along the Baffin Island coast generally indicates prevailing offshore (northerly and northwesterly) winds. After passing Big Island the *Emerald* made steady westward progress in open ice until, by the seventh week, it again became enclosed in ice near Nottingham Island. This fact implies that as late as early September, closed ice was moving south out of Foxe Basin and congesting the western entrance to the Strait. The vigorous southward movement of ice from Davis Strait in the east and from Foxe Basin in the west, as well as the open water along the Baffin Island coast, were probably indicators of prevailing northerly or northwesterly winds (Catchpole and Faurer 1983).

After sailing into Hudson Bay on September 6, the *Emerald* remained in open water until it encountered open ice on September 13 in the latitude of the Belcher Islands. These conditions persisted throughout the next 400 km until open water was again reached on September 19 at 54°N latitude. Apparently this ice at the southern end of the Bay lingered for at least another month, since it was again encountered by the *Emerald*, during its return voyage, at 57°N latitude on October 8 (Figure 6). Comments in the log-book indicate that in mid-September large bodies of ice remained in the vicinity of Cape Henrietta Maria. Thus, on September 17 the ship "fell in with close heavy Ice...a great deal of Ice..." (Hudson's Bay Company, log-book of Brig *Emerald*, C1/324, fol. 59, 1816) while sailing 50 km to the east of the Cape (Figure 8).

In the absence of historical records of summer sea-ice severity in Hudson Bay, the significance of the ice conditions described in the log-book of the *Emerald* must be judged against the yardstick of modern records. This is unfortunate, since these records span a brief period of 15 years from 1964 to 1979 (Sowden and Geddes 1980). The choropleths on Figure 7 illustrate the latest dates on which ice has been observed at locations in the Bay in the period of record. The tendency for ice to clear first in the area to the south of Southampton Island and also at the southern end of James Bay is apparent in the spatial patterns displayed (Figure 7). Likewise, these patterns reveal the tendency for last remnants of ice to persist in the area west of Cape Henrietta Maria.

Exceptional variations in the dates in Figure 7 occur in the area between the southern end of James Bay, Cape Henrietta Maria and the Belcher Islands. Within a distance of only 400 km, the earliest and the latest dates of last remnants of ice in Hudson Bay are encountered. This feature militates against the objective interpretation of the significance of the ice conditions described in the log-book of the *Emerald* in 1816. During its southward passage between Cape Henrietta Maria and the Belcher Islands, the *Emerald* encountered ice in mid-September in an area where it has not been observed after August 13 in the modern period. However, ice has been recently observed as late as September 10 in the area immediately to the west of this route. Two features of the ice

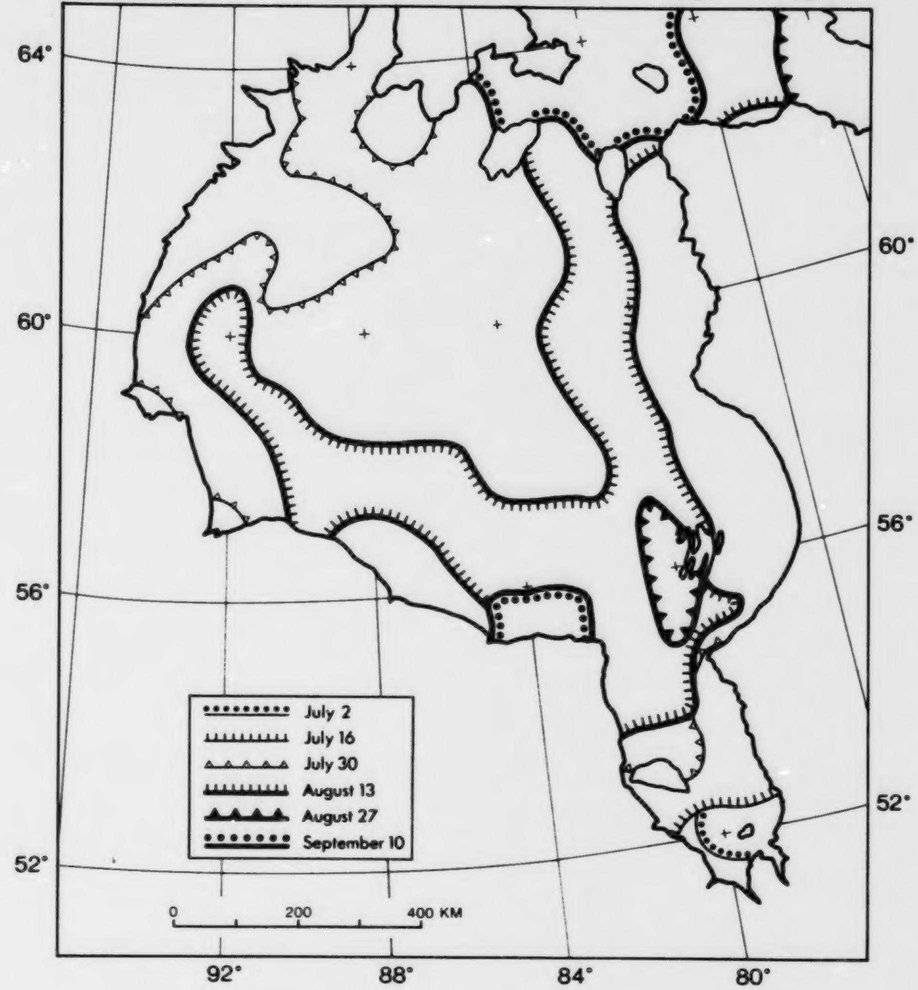
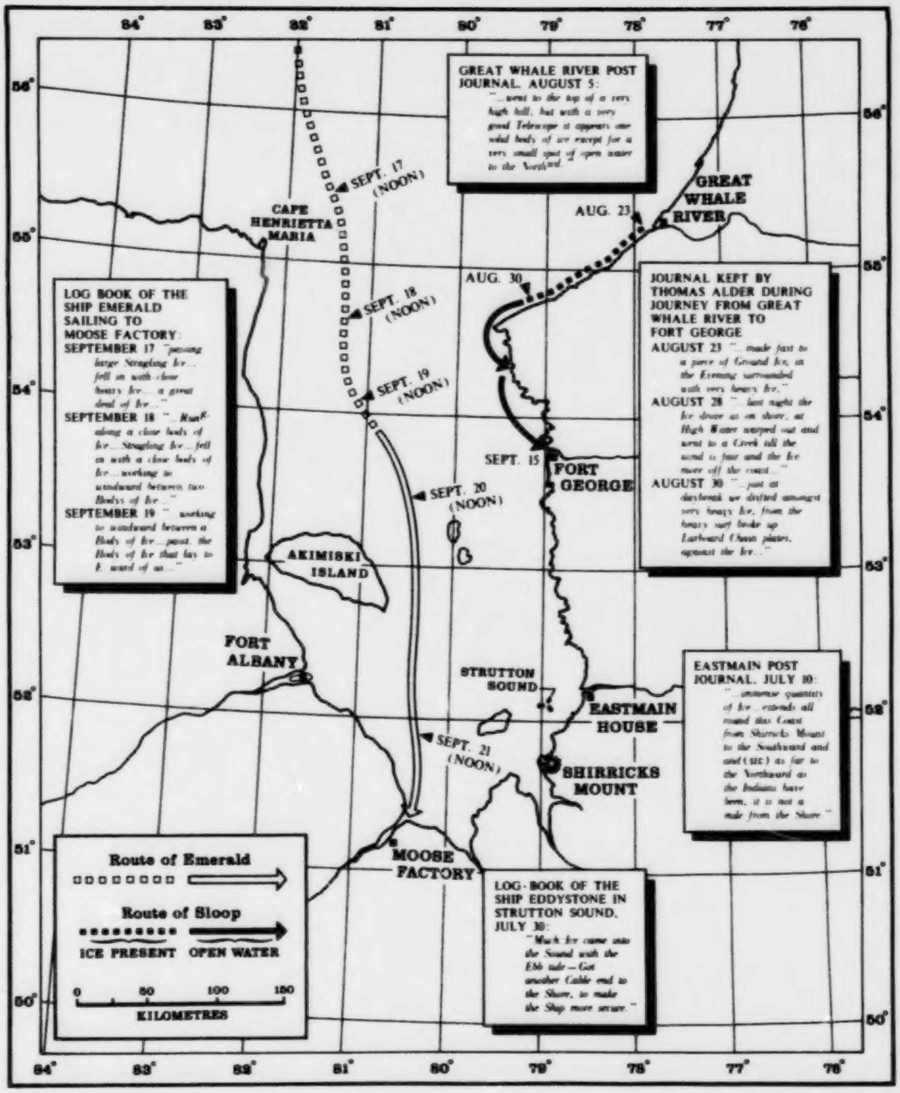

96°
92°
88°
84°
80°
76°
72°
64°
60°
56°
52°
60°
56°
52°
July 2
July 16
July 30
August 13
August 27
September 10
0
200
400 KM
92°
88°
84°
80°

FIGURE 7: *Maximum extent of ice in Hudson Bay observed at two-week intervals during late summer, 1964 to 1979. Maximum extent is defined as the area where, on the given date, ice has existed at least once during the period of observation. The symbols on the lines point towards the area of ice (after Sowden and Geddes 1980).*

distribution in 1816 appear to represent exceptionally late occurrences. The first includes the ice observed on September 18 and 19 to the south of 55°N latitude in James Bay, and the second involves the ice encountered in the latitude of the Belcher Islands on October 8 during the northward passage of the *Emerald*. In each of these cases the ice occurrence in 1816 was *over one month later* than the modern date of last ice.

Figure 8 presents ice descriptions from several other sources in the Hudson's Bay Company records. Together these indicate that, in the summer of 1816, late ice was widespread in James Bay and not restricted to the route followed by the *Emerald*. On July 10, the Eastmain post journal (Hudson's Bay Company, Eastmain journal, B59/a/96, fol. 2, 1816) referred to "immense" quantities of ice extending north along the coast from Shirricks Mount. This condition was not exceptionally late since it occurred in an area where ice has recently been observed as late as July 2. However, in 1816 the same area produced the conditions under which the ship *Eddystone* was threatened by 'much ice' while anchored in Strutton Sound on July 30 (Hudson's Bay Company, log-book of Ship *Eddystone*, C1/302, fol. 82, 1816). The Great Whale River post journal reported that, on August 5, the area of Hudson Bay visible by telescope from the top of a 'very high hill' was virtually completely ice covered (Hudson's Bay Company, Whale River journal, B372/a/3, fol. 27, 1816). The same journal later described ice conditions encontered by Thomas Alder when he sailed in a coastal sloop from Great Whale River to Fort George in late August. 'Very heavy ice' was encountered both on August 23 and August 30 (Hudson's Bay Company, Whale River journal B372/a/3, fol. 29 and 30, 1816). These descriptions are from an area where the latest modern ice observations are on July 30, though it is noteworthy that ice has recently been observed as late as August 27 within 50 km of this shore.

Hudson Strait provided clear evidence that ice conditions in the summer of 1816 were indicative of prevailing northerly or northwesterly winds. Likewise, in Hudson Bay several pieces of evidence point to exceptionally late ice at the southern end of the Bay and in James Bay. This finding is consistent with the advective and frictional affects of arctic air moving rapidly southwards.



84°
83°
82°
81°
80°
79°
78°
77°
76°
56°
55°
54°
53°
52°
51°
50°
SEPT. 17 (NOON)
CAPE HENRIETTA MARIA
SEPT. 18 (NOON)
SEPT. 19 (NOON)
SEPT. 20 (NOON)
AKIMISKI ISLAND
FORT ALBANY
SEPT. 21 (NOON)
MOOSE FACTORY
GREAT WHALE RIVER
AUG. 23
AUG. 30
SEPT. 15
FORT GEORGE
STRUTTON SOUND
EASTMAIN HOUSE
SHIRRICKS MOUNT
Route of Emerald
Route of Sloop
ICE PRESENT
OPEN WATER
0
50
100
150
KILOMETRES

GREAT WHALE RIVER POST JOURNAL, AUGUST 5:
"...went to the top of a very high hill, but with a very good Telescope it appears one solid body of ice except for a very small spot of open water to the Northward."

LOG BOOK OF THE SHIP EMERALD SAILING TO MOOSE FACTORY:
SEPTEMBER 17 "passing large Straggling Ice... fell in with close heavy Ice... a great deal of Ice..."
SEPTEMBER 18 "...Runᵍ along a close body of Ice... Straggling Ice... fell in with a close body of Ice... working to windward between two Bodys of Ice..."
SEPTEMBER 19 "...working to windward between a Body of Ice...past. the Body of Ice that lay to E. ward of us..."

JOURNAL KEPT BY THOMAS ALDER DURING JOURNEY FROM GREAT WHALE RIVER TO FORT GEORGE
AUGUST 23 "...made fast to a piece of Ground Ice, in the Evening surrounded with very heavy Ice."
AUGUST 28 "...last night the Ice drove us on shore, at High Water warped out and went to a Creek till the wind is fair and the Ice move off the coast..."
AUGUST 30 "...just at daybreak we drifted amongst very heavy Ice, from the heavy surf broke up Larboard Chain plates, against the Ice..."

EASTMAIN POST JOURNAL, JULY 10:
"...immense quantity of Ice...extends all round this Coast from Shirricks Mount to the Southward and and (sic) as far to the Northward as the Indians have been, it is not a mile from the Shore."

LOG-BOOK OF THE SHIP EDDYSTONE IN STRUTTON SOUND, JULY 30:
"Much Ice came into the Sound with the Ebb tide — Got another Cable end to the Shore, to make the Ship more secure."

FIGURE 8: *Descriptions of sea ice in late summer, 1816, contained in the Hudson's Bay Company records. [Sources: Hudson's Bay Company, log-book of brig* Emerald, *C1/324, fol. 59, 1816; Hudson's Bay Company, log-book of ship* Eddystone, *C1/302, fol. 82, 1816; Hudson's Bay Company, Whale River journal, B372/a/3, fol. 27, fol. 29-30, 1816; Hudson's Bay Company, Eastmain journal, B59/a/96, fol. 2, 1816].*

SUMMARY

1816 has been described as "the year without a summer," in the light of evidence from Europe and North America. These weather conditions are commonly, though not universally, attributed to the effects of stratospheric aerosols generated by the eruption of Mount Tambora in 1815. This paper presents a variety of evidence of weather conditions in the Hudson Bay region in 1816. In particular, it indicates the prevalence of meridional atmospheric circulation using evidence of sea-ice conditions in Hudson Bay and Hudson Strait.

ACKNOWLEDGEMENTS

I am indebted to Ms. Janet Dickie, Ms. Marcia Faurer and Mr. Michael Pidwirny, graduate students at The University of Manitoba, who assisted me in this research. The encouragement and stimulation that I have received from Dr. C.R. Harington, Paleobiology Division, National Museum of Natural Sciences, National Museums of Canada, Ottawa is gratefully acknowledged. I thank Mrs. Shirlee Smith, Hudson's Bay Company Archivist for the advice and assistance that she has extended to me over a period of several years. This research was supported by a contract received from the National Museum of Natural Sciences Climatic Change Project. Permission to consult and quote from its archives was given by the Hudson's Bay Company.

REFERENCES

Barry, R.G. 1981. The nature and origin of climatic fluctuations in northeastern North America. Géographie physique et Quaternaire 35:41-47.

Catchpole, A.J.W. 1980. Historical evidence of climatic change in western and northern Canada. In: Climatic Change in Canada. Edited by: C.R. Harington. Syllogeus No. 26:17-60.

Catchpole, A.J.W., and D.W. Moodie. 1978. Archives and the environmental scientist. Archivaria 6:113-136.

Catchpole, A.J.W., and T.F. Ball. 1981. Analysis of historical evidence of climatic change in western and northern Canada. In: Climatic Change in Canada 2. Edited by: C.R. Harington. Syllogeus No. 33:48-96.

Catchpole, A.J.W., and Marcia-Anne Faurer. 1983. Summer sea ice severity in Hudson Strait, 1751-1870. Climatic Change 5:115-139.

Crane, R.G. 1978. Seasonal variations of sea ice extent in the Davis Strait - Labrador Sea area and relationships with synoptic-scale atmospheric circulation. Arctic 31:434-447.

Danielson, E.W. 1971. Hudson Bay ice conditions. Arctic 24:90-107.

Dey, B. 1980. Seasonal and annual variations in ice cover in Baffin Bay and northern Davis Strait. Canadian Geographer 24:368-384.

Hammer, C.U., H.B. Clausen, and W. Dansgaard. 1980. Greenland ice sheet evidence of post-glacial volcanism and its climatic impact. Nature 288:230-235.

Hoyt, J.B. 1958. The cold summer of 1816. Annals of the Association of American Geographers 48:118-131.

Jacobs, J.D., and J.P. Newell. 1979. Recent year-to-year variations in seasonal temperatures and sea ice conditions in the eastern Canadian Arctic. Arctic 32:345-354.

Keen, R.A. 1978. The response of Baffin Bay ice conditions to changes in atmospheric circulation patterns. In: Proceedings of the Fourth International Conference on Port and Ocean Engineering under Arctic Conditions. Edited by: D.E. Muggeridge. Volume II. Memorial University, St. John's. pp. 963-971.

Lamb, H.H., and A.I. Johnson. 1966. Secular variations of the atmospheric circulation since 1750. Meteorological Office (London), Geophysical Memoirs 110. p. 57.

Landsberg, H.E., C.S. Yu, and L. Huang. 1968. Preliminary reconstruction of a long time series of climatic data for the eastern United States. University of Maryland, Institute for Fluid Dynamics and Applied Mathematics, Technical Note BN-571:1-75.

Landsberg, H.E., and J.M. Albert. 1974. The summer of 1816 and volcanism. Weatherwise 27:63-66.

Manley, G. 1974. Central England temperatures: monthly means 1659-1973. Quarterly Journal of the Royal Meteorological Society 100:389-405.

Markham, W.E. 1981. Ice atlas: Canadian arctic waterways. Canada, Atmospheric Environment Service. 198 pp.

Maxwell, J.B. 1982. The climate of the Canadian Arctic Islands and adjacent waters, Volume 2. Canada, Atmospheric Environment Service Climatological Studies 30:1-589.

Moodie, D.W. 1978. Gardening on Hudson Bay: the first century. The Beaver 309(1):54-59.

Newhall, C.G., and S. Self. 1982. The volcanic explosivity index: an estimate of explosive magnitude for historical volcanism. Journal of Geophysical Research 87:1231-1238.

Parker, M.L., L.A. Jozsa, Sandra G. Johnson, and Paul A. Bramhall. 1981. Dendrochronological studies on the coasts of James Bay and Hudson Bay (Parts 1 and 2). In: Climatic Change in Canada 2. Edited by: C.R. Harington. Syllogeus No. 33:129-188.

Parker, M.L., Paul A. Bramhall, and Sandra G. Johnson. 1982. Tree-ring dating of driftwood from raised beaches on the Hudson Bay coast. Report prepared for Canadian Forestry Service, Ottawa. p. 41.

Post, J.D. 1977. The last great subsistence crisis in the western world. The Johns Hopkins University press, Baltimore. 240 pp.

Schneider, Stephen H. 1983. Volcanic dust veils and climate: how clear is the connection? Climatic Change 5:111-113.

Sowden, W.J., and F.E. Geddes. 1980. Weekly median and extreme ice edges for eastern Canadian seaboard and Hudson Bay. Canada, Ice Climatology and Applications Division, Ottawa.

Stommel, H., and E. Stommel. 1979. The year without a summer. Scientific American 240:176-186.

Thompson, L.G., and E. Mosley-Thompson. 1981. Temporal variability of microparticle properties in polar ice sheets. Journal of Volcanology and Geothermal Research 11:11-28.

Walsh, J.E., and C.M. Johnson. 1979. An analysis of arctic sea ice fluctuations, 1953-77. Journal of Physical Oceanography 9:580-591.

Wilson, C.V. 1982. The summer season along the east cost of Hudson Bay during the nineteenth century, Part 1: General introduction; climatic controls; calibration of the instrumental temperature data, 1814 to 1821. Canadian Climate Centre. Report No. 82-4:1-223.

CORRECTIONS: ➡

Page 149. Lines 2-4. Delete: "While this rare event ... in the nineteenth century." Add: "During the nineteenth century, ships had to winter in the Bay on 12 occasions, five of these between 1811 and 1819."

Footnote. Add: "In 1819, The Prince of Wales and Eddystone returned safely to England, but the Weir, leaving Moose Factory nearly two weeks later than the Eddystone, on September 30, was unable to exit through the ice."

Page 160. Last paragraph. Line 4. Should read "... from (0° to <5°C) to (5° to <10°C)."

Last paragraph. Line 8. Should read "... the warm 1968 and colder 1969 seasons ..."

Page 182. First paragraph. Line 12. Should read "... to submarine crustal activity ..."

Page 186. Add:

Manley, G. 1957. Climatic fluctuations and fuel requirements. Scottish Geographical Magazine 73(1):19-28.

_____. 1978. Variations in the frequency of snowfalls in East-central Scotland, 1708-1975. Meteorological Magazine 107:1-16.

THE LITTLE ICE AGE ON EASTERN HUDSON/JAMES BAY: THE SUMMER WEATHER AND CLIMATE AT GREAT WHALE, FORT GEORGE AND EASTMAIN, 1814 TO 1821, AS DERIVED FROM HUDSON'S BAY COMPANY RECORDS

Cynthia Wilson[1]

INTRODUCTION

On August 5 1816, Thomas Alder, Master of the Hudson's Bay Company (HBC) Post at Whale River (Figure 1), climbed to the top of a very high hill to look out over the Bay through a good telescope. Except for a very small spot of open water to the north, the surface appeared as one solid body of ice. At the middle of the month, the ice was still closely compacted along the coast. Neither the sloop nor longboat could put to sea. The Indians had to leave even their canoes behind and walk with their furs along the shore to reach the Post. August 13 saw fresh snow after a scant five-week remission, and on the morning of the 18th the hills were "white over", with further snowfall during the day. Several old Indians said "they never knew so backward a season" (HBC, B372/a/3). Finding himself in a barren environment, with his men starving, sick with the scurvy and exhausted by the cold, and with no life-line, Alder in desperation and in defiance of Company policy decided to evacuate the Post - to try to fight a way back south through the ice. The sloop and longboat finally forced a way into the Bay on August 26. After a nightmare journey along the treacherous, rocky coast, through straggling and heavy ice with gales, freezing temperatures, blizzards and heavy rain ("...though I've been 20 years coasting along the different shores, I never knew a more miraculous escape" --- ibid.), they eventually rounded Cape Jones, to reach the comparative safety of Big River on September 15. Whale River Post was not to be re-opened until 1856.

The exceptionally cold summer of 1816 along the east coast of Hudson/James Bay was followed by another in 1817 ("...if summers I may call them ...", Alder: HBC B77/e/1a) before the weather at this season returned to within normal expectation for the early nineteenth century. In fact, unusually severe weather had occurred consecutively through winter and summer, with little relief, from the autumn of 1815 to late winter of 1818.

---

[1] P.O. Box 887, Station B, Ottawa, Ontario K1P 5P9
This study was carried out under contract to the Atmospheric Environment Service of Canada, and with the kind permission of the Hudson's Bay Company.

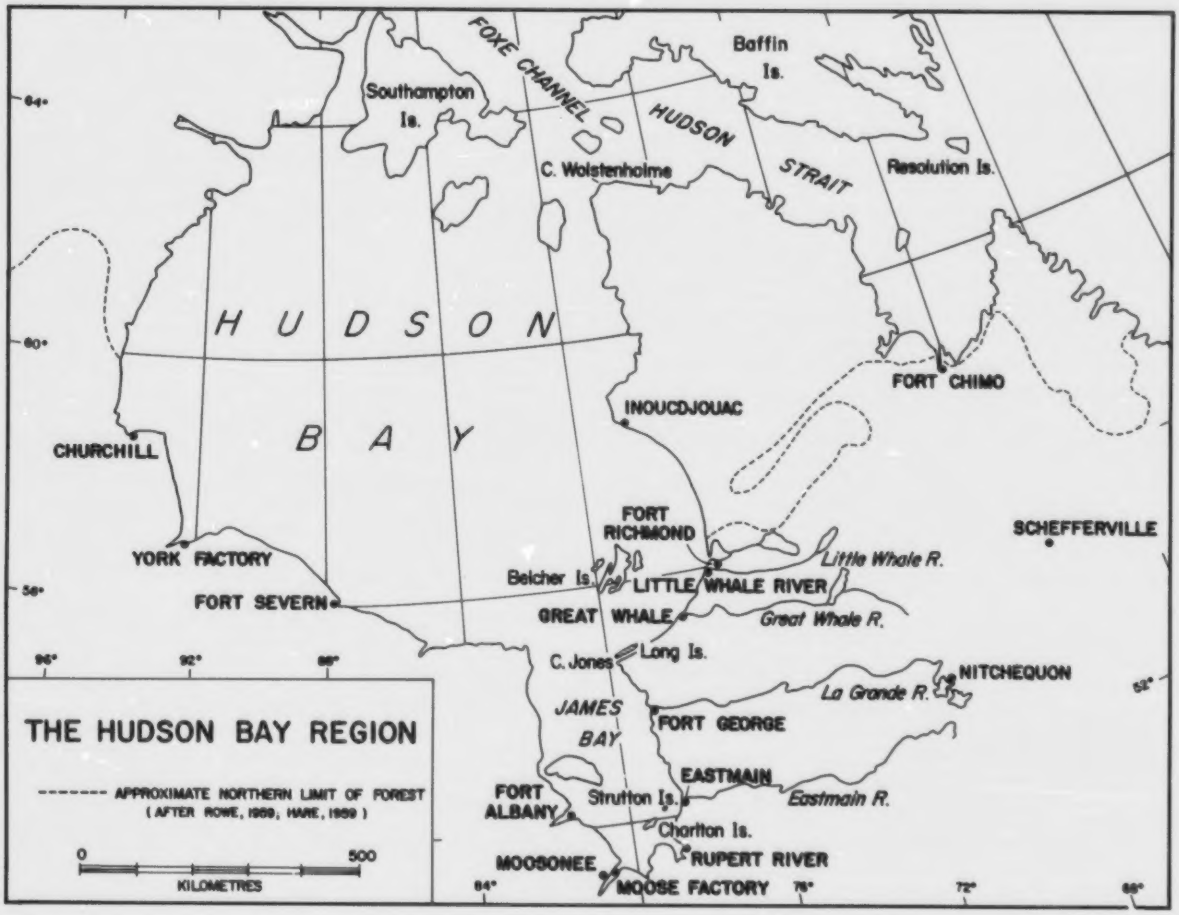

Southampton Is.
FOXE CHANNEL
Baffin Is.
HUDSON STRAIT
C. Wolstenholme
Resolution Is.
64°
60°
56°
52°
HUDSON BAY
FORT CHIMO
INOUCDJOUAC
CHURCHILL
SCHEFFERVILLE
FORT RICHMOND
YORK FACTORY
Belcher Is.
Little Whale R.
LITTLE WHALE RIVER
FORT SEVERN
GREAT WHALE
Great Whale R.
C. Jones
Long Is.
NITCHEQUON
La Grande R.
96°
92°
88°
84°
76°
72°
68°
JAMES BAY
FORT GEORGE
EASTMAIN
Eastmain R.
FORT ALBANY
Strutton Is.
Charlton Is.
RUPERT RIVER
MOOSONEE
MOOSE FACTORY
THE HUDSON BAY REGION
------- APPROXIMATE NORTHERN LIMIT OF FOREST (AFTER ROWE, 1959; HARE, 1959)
0
500
KILOMETRES

FIGURE 1: *General map of the Hudson Bay region.* Please note: *In 1966, the settlement of Great Whale became known as Poste-de-la-Beleine, and Great Whale River as Grande rivière de la Baleine. More recently, the Inuit name Kuujjuaraapik has been introduced. The name Great Whale is retained in this study for continuity and simplicity. Similarly, the more familiar Cape Jones has been preferred in this context over the official Point Louis - XIV. For the nineteenth century Posts, the original Whale River (Great Whale) and Big River (Fort George) are used to denote the historical period.*

The early autumn closure of Hudson Strait by ice in 1815 and 1816 had forced the ships, which each summer linked England and the Bay Posts, to winter over[1] near Eastmain. While this rare event had also occurred in 1811/12, it was to happen only once again (1833/34) in the nineteenth century. At a stage when the majority of the inhabitants lived very close to the land in territory marginal for life and unforgiving at the best of times, this relentless run of bad weather served to compound the impact of the 1816, 1817 seasons and to produce "almost unheard of conditions of poverty and distress" (HBC, B77/e/3).

An analysis of the carefully observed, detailed descriptions of the weather from 1814 to 1821 for Whale River, Big River and Eastmain (Figure 1) contained in the Hudson's Bay Company records, indicates that climate in the active season (May to October) 1816, 1817 was not only colder than that presently on record (from 1916) along the eastern region of the Bay, but that the situation in 1816 came marginally close to producing a residual snowcover. Indeed, in studying the weather from the autumn of 1815 to late winter 1818, there is a distinct sense of approaching a critical set of conditions, which had they persisted, might have led to the inception of permanent snowfields in New Québec/Labrador.

The years 1814 to 1821 include the volcanic eruption of Mount Tambora in Indonesia (April 1815) and span the second of a double cycle of low sunspot number - the lowest since the end of the Maunder Minimum, a century earlier.

This paper describes the main features of the summer climate along this coast at this time, and concludes with a brief discussion of three aspects of the results:

1. Summers 1815 to 1820, volcanic aerosol and sunspots.
2. Autumn 1815 to spring 1818, and the inception of permanent snowfields over New Québec/Labrador.
3. The social and economic impact of these seasons.

CONTEXT

There is already much evidence for the northeast United States and western and central Europe pointing to the summer of 1816 and the summer period 1811 to 1820 as the coldest of the nineteenth and twentieth centuries. In some cases this decade has been recognized as one of the coldest of the three centuries from 1550 to 1850, which are often referred to as the Little Ice Age. In the Hudson Bay/New Québec region, break-up dates for estuaries on the Bay, indicators of ice conditions in Hudson Strait and other proxy series for the west side of the Bay confirm the severe conditions at this time (Catchpole *et al.* 1976; Catchpole and Ball 1981). Further evidence is provided in the tree-ring series at Fort Chimo (Cropper and Fritts 1981) and Great Whale (Payette 1976; Parker *et al.* 1981), which also indicate an unusual degree of cold in the second decade of the nineteenth century, focusing on 1816, 1817. On the other hand, studies based on tree-ring evidence from northwestern North America (for example, Blasing and Fritts 1975) strongly suggest that the temperature anomalies during this period were out of phase with those in the northeastern States.

---

[1] In the autumn of 1817, the *Eddystone* left Moose Factory in mid-September and was able to return to England. The *Britannia* left York at the beginning of October and was forced back by heavy ice and bad weather. The ship was driven ashore at Severn on the way to winter at Charlton (Cooke and Holland 1978).

Although the nature, degree and extent of the influence of the Bay on Canadian climate have still to be fully assessed, the Hudson/James Bay region is believed to be particularly sensitive with respect to climate fluctuation. This vast sea spanning 10 degrees of latitude, open only to the north to Arctic waters and Arctic ice, extends the influence of polar climate into the heart of the continent in spring, and remains a cold source in summer. In late autumn, the presence of open water creates a snowbelt on the windward east coast. All forms of life are so finely tuned to climate along these marginal coastlands, that any unusually severe or prolonged anomaly can soon disturb the ecological balance and human life and activity.

To the north of Cape Jones (Figure 1), the coastal environment is bleaker. The Shield outcrops to form a low dissected plateau, much of it bare. Trees are generally confined to the less exposed sites and river valleys, with tundra occupying the narrow coastal terraces. To the south of the Cape, the Shield outcrop runs inland to the southeast. Here, the coastal landscape consists of low, rolling country of postglacial sands and clay, with open lichen woodland on the drier sites. Towards Eastmain, there is an increasing percentage of bog and swamp, and denser and more varied woodland growth. A marked climatic gradient occurs in the vicinity of Cape Jones; often the last remnants of the ice to clear the east coast of the Bay are found here.

DATA AND METHOD

In 1814, the Hudson's Bay Company initiated a program of weather observation, largely to study the possibilities of local agriculture at the various Posts (HBC, A6/18). At Whale River/Big River and Eastmain, Weather Registers were continued until the amalgamation with the North West Company in 1821. They contain fixed-hour observations of temperature, wind and weather, usually for morning, noon and evening hours, and at Whale River and Big River, daily maximum and minimum temperatures. These data, together with weather information in the Post Journals, Annual Reports, Correspondence and Ships' Logs have been used to construct a regional climatology for the summer seasons, May to October.

The observations of the various weather elements were checked as carefully as possible against modern standards, and tabulated so as to permit direct comparison with modern data sets for Great Whale, Fort George and Eastmain. Particular attention was given to the calibration of the historical temperature records; the procedures and assumptions have been reported in detail elsewhere (Wilson 1982, 1983A). With repect to the confidence limits of the resulting *corrected* monthly mean daily temperatures, statistically the error appears to to be less than ± 1°C from May to August, and within ± 0.5°C in autumn. Concerning the basic assumptions made, it is believed that any realistic changes would not alter the corrected values by more than about ± 0.5°C. Arising from these procedures, adjustments were also made to the mean daily maximum and mean noon readings. Taking into consideration the possible systematic and non-systematic errors which may be hidden in the modern temperature series at these weather stations, the limits of error seem acceptable for the historical series.

In comparing the climate of historical and modern periods, an analogue approach was taken. The modern analogues were defined in terms of the monthly mean daily temperature (i.e. extremely warm, cold or average months) at Great Whale, Fort George and Eastmain covering the period 1916 to 1976. The principal normal reference period is 1941-1970. For further information concerning method, and for a more detailed description of the weather and climate of 1814 to 1821, together with tables and other supporting evidence, see Wilson (1983B). An analysis of the present-day weather and climate, and a description of the coastal environment were included in the earlier report and paper (Wilson 1982, 1983A).

ASPECTS OF THE SUMMER CLIMATE, 1814 to 1821

Mean Daily Temperature Anomalies

Differences between the historical monthly mean daily temperatures for Whale River, Big River and Eastmain and modern reference values for Great Whale, Fort George, and Eastmain are illustrated in Figure 2. The shading and asterisks indicate where the historical mean was below the modern extreme on station record - from 1926 at Great Whale, 1916 at Fort George and 1960, Eastmain. The absolute temperatures are given in Table 1. It was a period of great fluctuation.

In general, the active season (May to October) was colder in the historical period along the east coast of Hudson/James Bay, and with the exception of May 1818 and June 1820, consistently so in the spring and autumn months.

The series of seasons is clearly divided into two distinct periods, indicating a major change of circumstances in 1818. Although the two summer months in 1815 were near or above normal, there was an unusually cold spring and fall, and the season closely resembled one of the coldest of recent times, 1969. The persistent severity in 1816 and 1817 was not only beyond that on modern record, but apparently greater than that experienced in the prior 20 to 30 years. In July 1816, the mean daily temperature at Whale River was nearly 6°C below the 1941-1970 normal, and just over 2°C below the lowest on record (1965). The modern standard deviation for July is 1.2°C. At Big River in 1817, every month was below the modern record, and exceptional cold was registered in May and October at both Big River and Eastmain; the season as a whole was 5°C below the present-day average.

In 1818, the climate suddenly "sprang back", with the warmth in May approaching the modern record for Fort George. The seasons 1818 to 1820 were milder, and the monthly mean daily temperature lay within the modern range of variability. Since 1818 was considered at the time to be remarkably mild and unusually favourable, this suggests that the average expectation at this time was lower than the present normal. The three exceptionally cold seasons were thus followed abruptly by three that were relatively warm for the early nineteenth century.

From autumn 1815 through April 1818, this coast experienced arctic conditions. In the consecutive summers 1816 and 1817, the arctic boundary (after Köppen) lay close to Eastmain - about 52°30'N, some 3.5 degrees of latitude to the south of its present average location near Richmond Gulf. The closest modern analogue is probably the 1965 season. The gradient

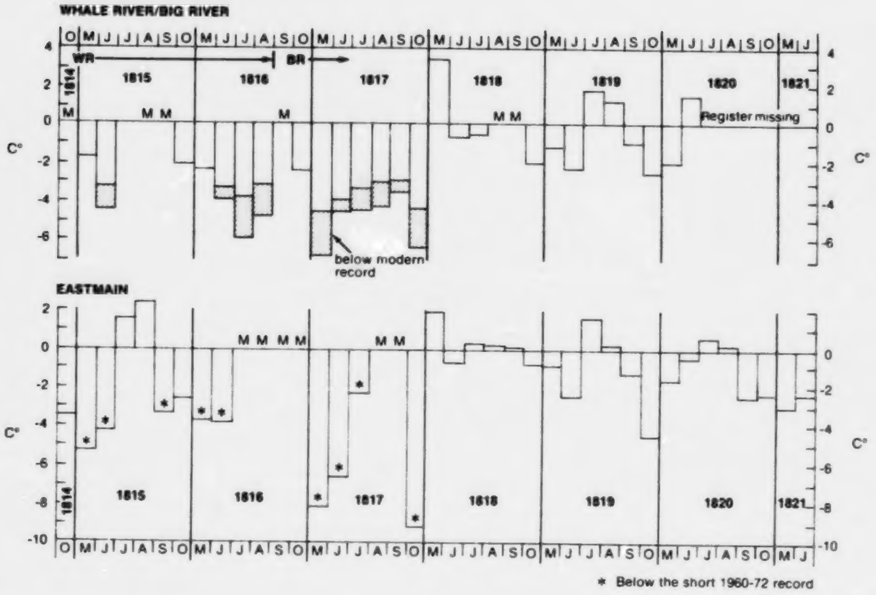


FIGURE 2: *Whale River, Big River, Eastmain, 1814-1821: mean daily temperature (adjusted values) expressed as* differences *from the 1941-1970 normals (WR, BR) or 1960-1972 averages (EM). The shading and asterisks indicate where the historical mean was below the extreme monthly mean on modern record.*

TABLE 1: *WHALE RIVER, BIG RIVER, EASTMAIN, 1814-1821: MEAN DAILY TEMPERATURE (ADJUSTED), MONTHLY AND SEASONAL VALUES (°C). BELOW: GREAT WHALE, FORT GEORGE AND EASTMAIN, MODERN NORMALS AND EXTREME MONTHLY MEANS.*[1]

YEARS	POST	M[3]	J	J[3]	A[3]	S[3]	O[3]	SP.	SU.[3]	AU.	SEASON[3]
1815	WHALE RIVER	-0.7^{26}	1.9	10.6^{26}	-	-	$(0.2)^{19}$	0.6	-	-	-
1816		-1.5	2.4	4.7	$(5.6)^{20}$		$(0.5)^{18}$	0.5	(5.2)	-	-
1817	BIG RIVER	-4.3	4.2	7.9	6.9	4.6^{29}	-3.4	-0.1	7.4	0.6	2.7
1818		5.9	8.1	11.8	-	-	0.9^{29}	7.0	-	-	-
1819		1.3	6.5	14.2	12.4	7.1	0.3	3.9	13.3	3.7	7.0
1820		0.5	10.4	-	-	-	-	5.5	-	-	-
1814	EASTMAIN	-	-	-	-	-	1.2	-	-	-	-
1815		-1.2	6.4	$(14.8)^{23}$	$(14.6)^{13}$	6.2	2.1	2.6	(14.7)	4.2	(7.2)
1816		0.3	6.7	-	-	-	-	3.5	-	-	-
1817		-4.1	3.9	10.9	-	-	$(-4.4)^{23}$	-0.1	-	-	-
1818		6.0	9.9	13.5	12.4	9.7	3.9	8.0	13.0	6.8	9.2
1819		3.2	8.0	14.8	12.4	8.4	0.2	5.6	13.6	4.3	7.8
1820		2.5	10.3	13.8	12.4	7.1	2.3	6.4	13.1	4.7	8.1
1821		1.1	8.3	-	-	-	-	4.7	-	-	-

	STATIONS	M	J	J	A	S	O	SP.	SU.	AU.	SEASON
1941-70	GREAT WHALE	0.9	6.3	10.6	10.3	7.5	2.3	3.6	10.5	4.9	6.3
S.D.[2]		(2.1)	(2.3)	(1.2)	(1.7)	(1.7)	(1.8)				
1941-70	FORT GEORGE	2.5	8.8	12.4	11.2	8.1	2.9	5.7	11.8	5.5	7.7
S.D.[2]		(2.1)	(1.5)	(0.9)	(0.9)	(0.8)	(1.2)				
1960-72	EASTMAIN	4.1	10.6	13.2	12.2	9.6	4.7	7.4	12.7	7.2	9.1
1951-80		5.1	11.3	13.7	13.0	9.2	3.8	8.2	13.4	6.5	9.4
S.D.[2]		(1.5)	(1.8)	(1.3)	(1.8)	(1.9)	(1.9)				

HIGHEST MEANS	M	J	J	A	S	O
Great Whale	4.8 (1960)	14.4 (1955)	14.9 (1931)	14.3 (1973)	13.2 (1968)	5.6 (1948)
Fort George	6.1 (1920)	12.6 (1930)	15.7 $\binom{1917}{1931}$	14.7 (1932)	10.4 (1940)	5.2 (1963)
Eastmain	6.9 (1968)	15.5 (1974)	16.0 (1974)	17.2 (1973)	14.2 (1968)	6.4 (1968)
LOWEST MEANS						
Great Whale	-3.8 (1956)	3.4 $\binom{1969}{1936}$	7.1 $\binom{1929}{1965}$	7.2 (1965)	4.2 (1969)	-1.3 (1936)
Fort George	-2.2 (1936)	4.9 (1936)	9.2 (1926)	8.2 (1934)	5.2 (1918)	-1.6 (1936)
Eastmain	2.2 $\binom{1969}{1970}$	7.9 (1961)	11.4 (1965)	9.6 (1965)	6.7 (1963)	0.9 (1974)

[1] Sources: Atmospheric Environment Service data tabulations, 1973, 1982.

[2] Standard deviation, 1951-80.

[3] Superscripts refer to number of days observed; bracketed numbers in this part of the table indicate more than five days are missing.

of air temperature, near the surface, normally steepens south of Cape Jones, and the anomalies appear to have been greater along James Bay than further north, especially in spring and autumn.

Bay Ice

The arctic summers 1816, 1817 were marked by unusually heavy ice and late break-up and melt in the east and south of the Bay. Moreover, *the 1816 season provides a marginal case for the carry-over of Bay ice from one year to the next*. Such a situation was considered exceptional for the period itself by both white and native peoples, and certainly has not been approached since regular ice observations began in the 1950s.

Although the relationship between the Bay surface conditions and weather is certainly not a simple one, even along this "windward" coast, the thermostatic effect of the Bay is sensed in the background temperature level during the warm season in any given year. When combined with winds from the Bay, the presence of ice or icy waters offshore late into the season is a major factor in the summer climate. Besides the persistent depression of air temperature near the surface arising from both the advection of cold air and the heavy, low cloud formations, there is the general lowering of the height of the freezing level, with its implications concerning summer and early autumn snowfall and the regional snowline. Colder summers tend to be characterized by a higher proportion of winds from the Bay. This was the case in 1816 and 1817.

In 1815, the eastern coast was apparently clear of ice relatively early in the season. However, ice was still present in mid-August in the northeast part of the Bay as far south as 58°30'N, 82°30'W, and the autumn freeze-up in the northeast was exceptionally advanced. On October 4, the ships ran into close heavy ice approaching Hudson Strait, and on October 10 found their exit blocked be a fast body of ice from shore to shore. The freeze-up was also unusually early in southeast James Bay where a great quantity of ice had already accumulated along the shore south of Eastmain by October 30.

The two seasons of severe ice conditions, 1816 and 1817[1], showed remarkably similar break-up patterns, with the last heavy ice compacted in the southeast of Hudson Bay extending into northern James Bay. The patterns of the final phase of wastage were not unlike the series of maximum ice/water limits for mid-August/mid-September drawn up by Sowden and Geddes (1980), based on grid-point data for the past 20 years. While the timing for 1817 was perhaps a week later than the modern maximum, particularly in James Bay, in 1816 these stages occurred some four weeks later, between mid-September and mid-October. The distribution and amount of ice in James Bay in mid-September 1816 was akin to that which would now normally be expected in mid-July. On October 14 1816, the passage into Hudson Strait was again found blocked, by both close, thick young ice and very high, heavy old ice. Given the cold autumn of 1816, it is most likely that the freeze-up again began early, as in 1815, so that the Bay was barely free of ice.

---

[1] For a note on recent seasons, see Appendix 1.

The break-up and melt in 1818 was relatively early in the east and southeast. In early July, Alder was able to sail around Cape Jones to Whale River, although some heavy, but open, ice was still present between the Cape and Little Whale River in mid-July. The passage was about three weeks earlier than Gladman's assessment of the average expectation, and about one week earlier than modern data would imply. After July 20, thre was no other mention of ice in the south or east. The clearing of the ice along this coast in 1819 was probably close to the average for the period, but in 1820 it was again early.

Mean Maximum, Minimum and Fixed-hour Temperatures

An idea of the mean diurnal temperature variation in the historical period, and a comparison with the modern record[1], is given in Tables 2 and 3.

As is the case today at this season, the anomalies for a given month are of similar sign for the mean daily maximum and minimum temperatures, with larger differences generally associated with the daily maximum - the changes in daytime heating. Tables 2 and 3 show a lowering of the maximum daily energy level of 5° to 6°C below the modern normal at Whale River in July 1816, and 3°C below the recent record in 1965. For the afternoon hours, the mean temperature reached only fractionally above 5°C. At Big River in 1817, there was a corresponding negative anomaly of more than 5°C in the daily maxima for at least five of the six months.

---

[1] Modern hourly data are only available for Great Whale; historical daily maximum and minimum readings were only taken at Whale River and Big River.

TABLE 2: *WHALE RIVER, BIG RIVER, 1815-1819: MEAN DAILY MAXIMUM (ADJUSTED) AND MINIMUM TEMPERATURES, THE ANOMALIES IN RELATION TO THE MODERN RECORD AND THE EXTREME DAILY VALUES (°C).*[1]

	WHALE RIVER / BIG RIVER							WHALE RIVER / BIG RIVER						
	M	J	J	A^{6}	S^{6}	O^{6}	YEARS	M	J	J	A^{6}	S^{6}	O^{6}	
MEAN DAILY MAXIMUM TEMPERATURE °C	-	-	-	-	-	(2.2)$^{\underline{19}}$	1815	-	-	-	-	-	(-2.0)$^{\underline{19}}$	MEAN DAILY MINIMUM TEMPERATURE °C
	3.1	7.2	9.1	(10.4)$^{\underline{20}}$		-	1816	-6.1	-2.3	0.4	(0.6)$^{\underline{20}}$		-	
	0.7	7.9	12.4	^{2}(12.4)$^{\underline{13}}$	^{2}6.8$^{\underline{25}}$	-1.6	1817	-9.3	0.5	3.5	^{2}(4.1)$^{\underline{13}}$	^{2}1.8$^{\underline{25}}$	-5.3	
	11.0	13.2	17.6	-	-	3.1$^{\underline{29}}$	1818	0.9	3.0	6.1	-	-	-1.3$^{\underline{29}}$	
	6.2	10.9	-	-	-	-	1819	-3.6	2.1	-	-	-	-	
DIFFERENCE FROM 1941-70 NORMAL	-	-	-	-	-	(-3.0)	1815	-	-	-	-	-	(-1.5)	DIFFERENCE FROM 1941-70 NORMAL
	-2.1	-4.1	-6.5	(-3.9)		-	1816	-2.6	-3.7	-5.2	(-5.8)		-	
	-6.8	-6.5	-5.3	(-3.3)	-5.1	-7.7	1817	-6.4	-2.6	-3.4	(-2.7)	-2.6	-5.5	
	+3.5	-1.2	-0.1	-	-	-3.0	1818	+3.8	-0.1	-0.8	-	-	-1.5	
	-1.3	-3.5	-	-	-	-	1819	-0.7	-1.0	-	-	-	-	
DIFFERENCE FROM LOWEST ON RECORD	-	-	-	-	-	-	1815	-	-	-	-	-	-	DIFFERENCE FROM LOWEST ON RECORD
	-	-0.1	-2.3	(-0.6)		-	1816	-	-0.8	-0.5^{3}	-2.2^{3}		-	
	-2.4	-2.2	-1.2	(0.0)	-1.8	-3.0	1817	-1.9	-	-1.2	-	0.0	-0.7	
EXTREME MONTHLY MAXIMUM (UNCORRECTED)	-	-	-	-	-	(6.1)$^{\underline{19}}$	1815	-	-	-	-	-	(-5.6)$^{\underline{19}}$	EXTREME MONTHLY MINIMUM (UNCORRECTED)
	20.0	26.1	25.6	(27.8)$^{\underline{20}}$		-	1816	-12.8	-6.7	-2.8^{5}	(-1.1)$^{\underline{20}}$		-	
	11.7	23.3	21.7	(22.8)$^{\underline{13}}$	17.8$^{\underline{25}}$	7.2	1817	-23.9^{5}	-6.1	-2.8	(0.0)$^{\underline{13}}$	-1.1$^{\underline{25}}$	-12.2	
	32.2^{4}	27.2	29.4	-	-	15.6$^{\underline{29}}$	1818	-8.9	-1.1	1.1	-	-	-11.1$^{\underline{29}}$	
	23.9	25.0	-	-	-	-	1819	-17.8	-3.3	-	-	-	-	

[1] Sources: Atmospheric Environment Service data tabulations, 1973.

[2] These values were not used to obtain the mean daily temperature.

[3] Adjusted to account for modern changes in the climatological day.

[4] Extreme maximum 32.2, above modern record.

[5] Extreme minimum -2.8, -23.9, below modern record.

[6] Underlined superscripts refer to number of days observed; bracketed numbers indicate more than five days are missing.

TABLE 3: *WHALE RIVER, BIG RIVER, EASTMAIN, 1814-1821: MEAN HOURLY TEMPERATURES (^{o}C). (NOON VALUES, ADJUSTED).*

		WHALE RIVER / BIG RIVER							EASTMAIN^{2}			
M^{3}	J	J^{3}	A^{3}	S^{3}	O^{3}	Years	M	J	J^{3}	A^{3}	S	O^{3}
8 am							6 am					
-	-	-	-	-	-	1814	-	-	-	-	-	0.8
$-0.6^{\underline{26}}$	$^{1}2.0$	$10.8^{\underline{26}}$	-	-	$(-0.6)^{\underline{19}}$	1815	$-^{1}4.2$	3.2	$(10.3)^{\underline{23}}$	$(13.0)^{\underline{13}}$	4.4	0.6
-2.4	$^{1}2.5$	$^{1}4.8$	$^{1}(4.5)^{\underline{20}}$		$(0.2)^{\underline{18}}$	1816	-1.8	4.8	-	-	-	-
$-^{1}4.6$	4.1	7.2	$^{1}5.3$	$^{1}3.2^{\underline{29}}$	$-^{1}4.3$	1817	$-^{1}5.7$	1.9	8.4	-	-	$-^{1}5.6^{\underline{23}}$
6.8	7.9	9.8	-	-	$0.4^{\underline{29}}$	1818	4.1	8.3	11.5	10.2	7.1	2.3
1.6	5.4	13.3	9.1	5.4	-1.0	1819	0.9	6.3	13.1	10.1	6.9	-0.9
-1.4	10.2	-	-	-	-	1820	1.1	9.1	11.6	10.6	5.3	0.7
-	-	-	-	-	-	1821	-0.8	6.6	-	-	-	-
Noon							Noon					
-	-	-	-	-	-	1814	-	-	-	-	-	2.2
-	-	-	-	-	$(1.1)^{\underline{19}}$	1815	3.7	8.7	$(16.5)^{\underline{23}}$	$(16.4)^{\underline{13}}$	8.3	4.9
0.4	$^{1}4.1$	$^{1}5.7$	$^{1}(7.6)^{\underline{20}}$		$(0.4)^{\underline{17}}$	1816	0.4	7.6	-	-	-	-
$-^{1}1.7$	5.3	10.1	$^{1}7.7$	$^{1}5.8^{\underline{29}}$	$-^{1}2.7$	1817	-0.4	6.9	12.1	-	-	$^{1}(-2.0)^{\underline{23}}$
8.1	9.7	14.2	-	-	$1.9^{\underline{29}}$	1818	7.7	12.4	16.7	15.8	12.1	6.7
3.4	8.4	-	-	-	-	1819	4.6	10.1	16.8	14.7	11.0	2.0
-	-	-	-	-	-	1820	3.4	12.1	15.8	14.9	9.2	3.9
-	-	-	-	-	-	1821	2.9	10.1	-	-	-	-
5 pm							6 pm					
-	-	-	-	-	-	1814	-	-	-	-	-	1.0
$0.3^{\underline{26}}$	$^{1}3.2$	$12.1^{\underline{26}}$	-	-	$(0.9)^{\underline{19}}$	1815	0.8	8.4	$(18.1)^{\underline{23}}$	$(15.1)^{\underline{13}}$	6.9	2.9
-0.1	$^{1}3.2$	$^{1}5.8$	$^{1}(4.8)^{\underline{20}}$		$(0.4)^{\underline{18}}$	1816	1.3	7.4	-	-	-	-
$-^{1}3.6$	4.7	$^{1}9.1$	$^{1}7.7$	$5.9^{\underline{29}}$	$-^{1}3.7$	1817	$-^{1}3.3$	4.8	12.2	-	-	$^{1}(-3.8)^{\underline{23}}$
8.7	10.2	15.0	-	-	$1.3^{\underline{29}}$	1818	6.9	10.3	14.2	13.7	11.3	4.8
3.9	8.0	16.8	17.5	8.8	1.2	1819	4.5	8.5	15.3	13.7	9.1	0.7
3.1	12.0	-	-	-	-	1820	2.9	10.4	14.8	13.2	7.9	3.3
-	-	-	-	-	-	1821	2.0	8.8	-	-	-	-

---

[1] Values below those on modern record for Great Whale.

[2] At Eastmain, 1814 to 1816, the morning and evening observations have been reduced to 6 am, 6 pm (Wilson 1982, 1983B).

[3] Underlined superscripts refer to number of days observed; bracketed numbers indicate more than five days are missing.

The largest depression of the daily minimum temperatures (more than 5°C below normal) occurred in July and August 1816[1] (WR) and May and October 1817 (BR). August 1816 thus provides an interesting exception (and contrast to July), in that the negative anomaly in the minimum (some 2°C beyond the present extreme) was considerably greater than that for the maximum. This is also reflected in the mean deviations at 8 am and 5 pm compared with that at noon. The midday temperature in August 1816 was in fact close to that of 1965. These differences between the degree of daytime heating in July and August 1816 appear to be real, and brought about by a change in the synoptic weather patterns at the end of July, and by rapid local changes in temperature and weather characteristic of this coast, especially with ice offshore. For example, on August 3 Alder writes "...at 11 am 82°F, in the evening at 5 pm 32°F" (HBC, B372/a/3). From the Weather Register for August, it appears that cloud, precipitation and thick fog were more frequent in the early evening than at midday. The proximity of the mean daily minimum to 0°C in August 1816 suggests the influence of the ice on the night-hour temperature (see below).

For Big River and Eastmain, it is useful to note those hourly means that are below the present Great Whale record, as this implies an important negative anomaly for those more southerly locations. Here, 1817 is outstanding - in particular the remarkably cold May and October.

Monthly Temperature Frequencies

The monthly frequency distributions for the daily minimum and the morning and early evening temperatures[2] in the early nineteenth century were remarkably similar in form and extreme values to those for corresponding cold, warm or average months today (Wilson 1983B). This suggests how far quite large climate fluctuations can be accommodated by relatively subtle changes in weighting within the known, modern frequency distributions, without reaching out towards new limits. One or two examples serve by way of illustration, and suggest the degree of influence of the Bay surface.

---

[1] Observations from August 1 to 20 only.

[2] The individual daily maximum and noon readings for the historical period are not always directly comparable with the modern data, owing to instrument error and exposure, and are not included here. For a discussion of these frequencies see Wilson (1983B).

Compared with the two coldest midsummer months at Great Whale, July and August 1965, the big difference in the frequency of daily minimum values for Whale River, July 1816, was the increase in the number of days just below freezing by almost one week, at the expense of days $\geq 5°C$. For August, this was even more significant; in 1965, there were no days below freezing, against five in the first 20 days of August 1816. Summer frosts at Great Whale and Fort George are usually advection frosts related to the ice or icy waters offshore. By contrast to 1816, the coastal waters in August 1965 were open in the vicinity of Great Whale and over most of the Bay, and the presence of open water was enough to maintain temperatures just above this critical level. However, the interactions between offshore ice and coastal temperatures are not always clear. In August 1969, with heavy ice offshore, no minimum temperature at Great Whale fell below 1.7°C, and the mean daily temperature was 0.7°C above normal. In July 1816, the frequent presence of clear arctic air appears to have been a contributing factor. In 1818, the frequency pattern of minimum temperatures for the very warm May at Big River is mirrored by that for May 1960, one of the warmest on record at Fort George. At Big River in July 1818, the temperature did not fall below 0°C, and only one such day was reported in June; the break-up and melt of the Bay ice was early this season.

Comparing the 8 am and 5 pm temperature frequencies at Great Whale in recent years and at Whale River in 1815 and 1816, the control exerted by the state of the Bay surface in the two midsummer months is again evident with respect to the 0°C threshold. This is as cleanly defined in the historical as in the modern data. The sharp cut-off in the distribution at 0°C, even in July 1816, is impressive. In August 1816, the single observation below was only marginally so. In general, the extreme cold of June 1815 and through the 1816 season is seen in the greater clustering on either side of the zero rather than in any greater increase in frequency in lower ranges. This is also the case today.

Taking into consideration the difference in latitude and exposure, the 8 am and 5 pm frequencies for Big River, 1817 to 1820, again correspond closely to those for Great Whale today. The grouped May - July distributions for 1817 to 1819 indicate that a warmer season was marked then as today by a shift in the modal temperature from (0° to $<5°C$) to (0° to $<10°C$). It is interesting to note that where frequencies can be grouped for the entire May - October season, they are identical in form to those for the first half of the season. A critical analysis of daytime hourly temperatures and wind direction at Great Whale during the warm 1968 and 1969 seasons (Wilson 1982) strongly suggested that differences in frequencies between 0° and 10°C were closely tied to the Bay surface temperature, and that this upward shift in the mode is in large measure a direct and indirect response to early clearing of the ice and strong heating of the offshore waters. Although the latter in turn is often linked with more frequent and persistent winds from the land (usually coupled with warmer, sunnier weather along the coast), over the season as a whole, winds from the Bay prevail.

Temperature Thresholds

Where these seasonal fluctuations in temperature occur around thresholds critical with respect to many physical and biological processes (for example, 0°, 5°, 10°C), as is the case along this coast, the impact can be far-reaching. The modern mean daily temperature in midsummer averages from just above 10°C at Great Whale to about 13°C at Eastmain (Table 1). Any unexpected lowering of the seasonal temperatures can become crucial for growth and survival.

The Active Season - River Ice and Ground Frost

At Whale River and Eastmain in 1815 and 1816, the beginning of the *active season* (defined as the passage of the mean daily temperature through 0°C) was two to three weeks later than the modern normal, akin to that of 1972. At Big River and Eastmain in 1817, the start of the season was even tardier - about four to five weeks later than the present normal; it ended at Big River over three weeks early, to give an active season of about 122 days, nearly two months shorter than the 1941-1970 average. In 1817, river ice at Big River did not begin to break until June 16. On June 12, men had crossed the river on the ice "...a thing I suppose which has never occurred at so late a period" (Alder, HBC, B77/a/3). At Eastmain, Russell concurred: "...such a spring as the present was surely never known in this country. The 30th of May and the river still as strong as in winter" (HBC, B59/a/96). This view is supported by the series of break-up dates for Eastmain, from the late 1740s (Catchpole and Ball 1981), in which 1817 is the latest on record. The ice eventually broke on June 7. Alder also records the severity of the frost in the ground at Whale River and Big River, 1815 to 1817. On June 6 1816, he describes ..."breaking up the garden ground with pick axes, as it's at this late period as solid frozen as in winter" (HBC, B372/a/3). At Big River in 1817, the depth of the frost in the sandy terrace on May 23 was over 2m,

"...such a late season I suppose was never known..." (HBC, B77/a/3). Alder had lived along this coast for more than 20 years. *It is almost certain that "seasonal" ground ice remained in some places through to the following winter in 1816 and 1817.* This region lies today in the zone of discontinous permafrost.

After a harsh winter[1], spring arrived early in 1818. The passage through 0°C occurred about the third week in April at both Posts. The timing is comparable with that for the anomalously warm spring of 1968. The river ice first broke on May 13/14, May 23 saw the willows in leaf at Big River, and on May 24, temperatures soared at both Posts to around 30°C -- "...the hottest day I ever experienced in this country" (Gladman at Eastmain, HBC, B59/a/99). Although the end of the active season was similar (EM) to or earlier (BR) than the average date today, the length of the season at both Posts (193/181 days) was comparable to favourable years today. The 1819, 1820 seasons, while shorter than the modern average, were probably close to "normal" for the period.

The Growing Season - Frost

The timing and length of the general *growing season,* as defined by a mean daily temperature above 5°C, and of the height of the *summer season* (above 10°C), 1814 to 1821, showed similar year-by-year patterns as those for the active season as a whole. In this region today, the period favourable for growth can be seriously curtailed by late spring and early fall frosts. This was evident in the early nineteenth century, when such frosts were often very sharp or killing.

Critical at Whale River in 1816 (and probably 1817) was the near-obliteration of the growing season per se for many plants in exposed places, and at Big River in 1817 the absence of any sustained period above the 10°C threshold (Table 1). Frosts were reported throughout each of these summers at all three Posts. In 1815, the relatively mild midsummer was offset by an exceptionally cold, late spring.

By contrast, the very warm, early spring in 1818, analogous to that of 1968, heralded a long season with mean daily temperatures above 5°C, and a substantial core-period above

---

[1] A table of monthly mean winter temperatures (uncorrected), 1814 to 1821 is given as an appendix (Table 5).

10°C. Even then, the isolated but sharp frost at Big River on June 25 severely damaged the potatoes, and there was a hoar frost at Eastmain on August 18. To the north, Alder's report during a visit to Whale River indicates sharp frosts on July 10 and 14. Midsummer 1819 brought extended spells of very warm, dry weather to eastern James Bay, but again frosts occurred at both Posts as late as July 3-5, with a severe frost as early as August 8. At Big River, where the latter damaged the potatoes, Alder reported frequent sharp frosts during the growing season. The most favourable frost-free season was that of 1820 - 81 days at Big River and 78 at Eastmain, compared with the modern average duration of 71 and 44 days[1], respectively. For much of the vegetation, any lesser midsummer heat in 1818 and 1820 may have been compensated by the greater warmth of the spring and the length of the growing/frost-free seasons. During this decade of the nineteenth century, the cut-off in the autumn is noticeably earlier, even in favoured seasons.

The Heating Season

Through tolerance, custom, clothing and necessity, the outdoor temperature threshold value at which indoor heating was considered comfortable was certainly less than today - even for Masters and Officers. An analysis of the series of indoor/outdoor temperature readings taken by Joseph Colen at Fort York in 1796 (Royal Society, MA 172) suggests that regular fires were maintained in the living quarters during the day once the outdoor temperature was below about 11° or 12°C. (This is slightly lower than Manley's (1957) assessment for Central England - a mean daily temperature of 13.3°C). Using this threshold, there would at present normally be several weeks in midsummer when fires would not be required on a regular basis at Great Whale; this period lengthens to the south, and by Eastmain extends from about mid-June to the beginning of September. The hourly means for historical period (Table 3) indicate that in 1816 and 1817, fires would have been necessary most days through the season at Whale River, to meet this standard of comfort, and except for brief periods in July, at Big River and Eastmain. With the warmth of spring and early autumn in 1818, the period without fires was probably close to or a little longer than it would average today at Fort George and Eastmain, and was very favourable for the early nineteenth century.

---

[1] The modern observing site at Eastmain is on organic terrain and more susceptible to frosts than those at Big River/Fort George. There is evidence to suggest that the historical site at Eastmain was less sensitive.

Snowfall

Compared with today, the most remarkable features of the May-to-October precipitation from 1814 to 1821 (Table 3, Figure 3) are:

(1) A greater total number of days with snowfall[1], and markedly so at Big River and Eastmain;

(2) A higher frequency of snowfall in autumn, as well as in late winter/spring. In October, at all three Posts, there were many more days with snow than rain. It is the reverse today;

(3) The shortening of the snow-free period.

These differences are especially noticeable along James Bay, and are greatest in 1816 and 1817. In the historical period, the late autumn/early winter snowbelt was more marked with a southward extension of the core zone of maximum snowfall from Great Whale (as at present) towards Eastmain. In addition, its onset was earlier, and in 1817 (and 1819) there was a shift from an annual highest monthly frequency of snowfall in November forward to October.

In 1816 at Whale River, the number of days with snowfall, although close to the present-day maximum, was still within modern range. The signficant difference lay in July/August. There were not only more days with snow in July than on record, but *snow fell on 7 days in August*. August is a key month. Through the modern period, there are no reports of snowfall in August at Great Whale. Thus, 1816 is the only year so far on record with snowfall in every month. There is no information for August 1816 at Eastmain, but the 12 days with snow in May and the falls in June and July represent a major anomaly. During the 1817 season at Big River and Eastmain, the frequencies in May/June were again high and many of the falls heavy, but the midsummer months (July/August) appear to have been without snowfall.

---

[1] Although this may be due in part to the modern definition of measurable snowfall, many of the historical occurrences were reportedly heavy, so that the differences appear to be real.

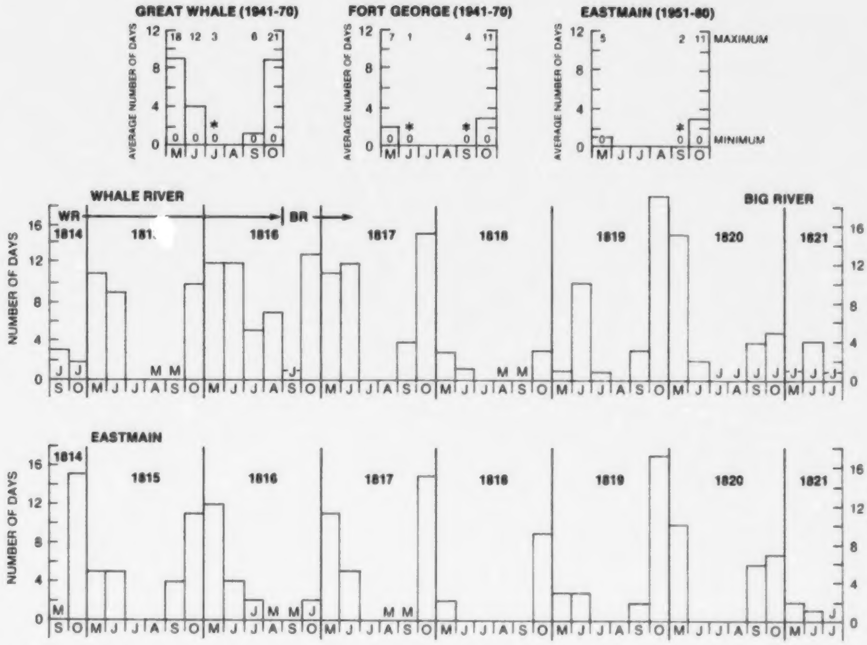


FIGURE 3: *Whale River, Big River, Eastmain, 1814-1821: number of days with snowfall, together with modern reference values for Great Whale, Fort George and Eastmain. (The climatological day beginning at 8 am at Whale River and Big River, 6 am at Eastmain.) M, data missing; J, Journal entries, no Register; *, less than 1 day.*

Along the east coast of James Bay, 1818 came closest to the modern normal with respect to snowfall. Even then three days with snow were reported in July at Whale River, on two of which the fall must have been heavy (see below). This would now be considered an extreme event. As is the case today, snow occurred principally with west to north winds from May to August, northwest and southeast in autumn.

Snowcover

Based on direct and indirect information contained in the Post Journals, I have attempted to reconstruct the nature and timing of the seasonal snowcover for the period of 1814 to 1821 (cf. Wilson 1983B). Apparently there was a much greater seasonal accumulation of snow along the James Bay coast in this decade than at present. Of particular interest is the period 1816 to 1818. There is little doubt that the presence of a late snowcover, whitened by frequent fresh snowfalls, was an exacerbating element in the severity of the 1816, 1817 seasons.

At Whale River, exceptionally heavy snow fell in the first half of April 1816[1]. On April 8, Alder wrote: "On Saturday last (6th) a heavy fall of snow and this day the same, not surpassed at any period during the winter; the aggregate depth of snow is impossible to form any idea of; the continual gales of wind which prevail here, drifting it in immense bodies in some places whilst in others the ground is barely covered;..." (HBC, B372/a/3). In early April, the snowcover south of the Post was so deep that the men had not been able to travel with the big sled. On April 15, the weather turned milder with little further snowfall until May 14, and it is believed that the windswept open terrace was mostly snow-free by early May. However, travel remained difficult south of the river owing to deep snow and drifts through at least mid-May. During the last half of May, in June and July, a record 29 days with fresh snow were reported at Whale River, with probably a renewed snowcover on the terrace between May 14 and 23. In June and July snow was said to cover the ground near the Post on eight mornings -- by as much as 5 cm on June 28, July 2 and 8, and 2.5 cm on July 9. Snow was reported again as early as August 13. On the morning of the

---

[1] A table showing the frequency of precipitation during the winter months is included as an appendix (Table 6).

18th, the hills were "white over"; more snow fell that day, with further heavy falls on August 22 and 24. (Alder evacuated the Post on August 26.) Given the amount of snow in the bush in spring, and the nature of the summer, *it seems most probable that snow lingered throughout the season in less-favoured sites in the hills, valleys and wooded areas.*

At Eastmain, the frequency of snowfall through the 1815/1816 winter was well in excess of the modern average. By late November/early December, the snowcover was already 0.75 m deep both on the plains and in the bush. The very cold winter and exceptionally heavy falls from mid-March to mid-April left large accumulations by early spring 1816. As at Whale River, a thaw set in, mid-April, which appears to have cleared much of the snow from the immediate vicinity of the Post by early May, and made travel difficult in the bush. But there were a further 16 days with snowfall through May and June (on May 9 a man was snowblind; on June 4 the men were throwing snow off the potato garden). There was probably a temporary snowcover on the terrace from May 13 to 23. And most remarkable for Eastmain, snow and sleet fell during most of the day on July 7 and 8, with a strong northwesterly gale. Although information is incomplete from August to mid-October, there is reason to believe that snowcover in many sites along eastern James Bay was prolonged into the summer season beyond modern experience.

In early spring 1817, the snowcover conditions appear to have been even more severe than in 1816, as there was no period of thaw. At Big River and Eastmain, daytime temperatures remained with few exceptions below 0°C until the end of May, while very heavy falls of snow occurred from mid-March until late May. Russell described the drifts as over 4 m high around Eastmain on May 31. It appears that the winter snowcover cleared from much of the terrace near the Posts during early June; again, considering the accumulation, it must have been very much later in the bush. But there were 12 days with snow in June at Big River until the final fall June 25, and five days at Eastmain, the latest June 23. The lateness of the snowcover is underlined by the entry in the Big River Journal for June 9. After noting that one man was incapable of duty owing to snowblindness, Alder adds: "Snowblindness this spring has been more prevalent than I ever knew before..." (HBC, B77/a/3). In autumn 1817, the first snow fell at Big River on September 12. Snow covered the ground from September 28-30, when it was reported 7.5 cm deep. During the abnormally cold October, heavy precipitation at both Posts fell almost without exception as snow, and the winter cover began on October 8 (BR) and October 9 (EM). Unlike 1816, the midsummer of

1817 was unusually wet along eastern James Bay (see below), and bouts of heavy or continuous rain most likely cleared much of the snow before the new season began. Given the sustained depression of seasonal temperature through 1817, it is a pity that there is no information this summer for the more northerly site, Whale River, where lower temperatures may have straddled the snow/rain threshold.

In contrast, by late winter 1818, snowcover was shallow, especially at Eastmain. Thaw occurred as early as March with the advent of southerly flow and some rain, and given the mild April, the cover was probably lost on the exposed terraces by early April, and on the marshes and swamps by the end of the month. At Big River, the clearance was some two to three weeks later. There was still occasional snowfall until May 12 at Eastmain, and as late as June 16 at Big River. This was considered so favourable a season, and yet Alder's temporary stay at Whale River in July revealed 10 cm of snow on the ground the morning of July 10. Although this was followed by heavy rains, a further 7.6 cm was reported July 14.

Rainfall

The largest positive rainfall anomalies in the early period were from July to September 1817 at Big River, 1820 at Eastmain, and July 1818 at both Posts. Drought occurred in 1819. In the 1817 season at Big River, there were 50% more days with precipitation than today's average (Table 4). The rains in August washed away numbers of young cabbage plants and together with repeated night frosts almost completely destroyed the potatoes and other roots. In September, rain proved disasterous for the haymaking.

Given the warmer summers of 1816 and 1820, the more frequent rainfall created fewer problems. However, the extremely wet weather at the end of August 1820 interfered with the haying at Eastmain. There was less rain in 1819 than the modern average, especially at Eastmain. Following the dry spring and early summer, gardens were withering by late July under extreme heat and drought, and the woods were on fire in all directions by early August. At Big River, May 1819 was almost without precipitation, and the woods to the north were reported burning in early June.

TABLE 4: *WHALE RIVER, BIG RIVER, EASTMAIN, 1814-1821: NUMBER OF DAYS WITH REPORTED PRECIPITATION, RAIN AND SNOW - THE CLIMATOLOGICAL DAY BEGINNING AT 8 AM (WR, BR) AND 6 AM (EM). BELOW: THE NORMAL PRECIPITATION AT GREAT WHALE, FORT GEORGE, EASTMAIN.*

		WHALE RIVER / BIG RIVER							EASTMAIN						
		M	J	J	A	S	O[2]	Total	M	J	J	A	S	O	Total
1814	PREC.	-	-	-	-	-	-	-	-	-	-	-	-	20	-
	RAIN	-	-	(0)	(1)	(7)	(1)	-	-	-	-	-	-	8	-
	SNOW[4]	-	-	(1)	(0)	(3)	(2)	-	-	-	-	-	-	15	-
1815	PREC.	17	17	$6^{\underline{27}}$	-	-	$14^{\underline{19}}$	-	8	12	$7^{\underline{23}}$	$7^{\underline{13}}$	11	18	-
	RAIN	8	9	6	-	-	6	-	3	8	7	7	9	9	-
	SNOW[4]	11	9	0	-	-	10	-	5	5	0	0	4	11	-
1816	PREC.	15	15	12	$15^{\underline{25}}$	-	17	-	15	14	-	-	-	-	-
	RAIN	3	8	8	8	(4)	6	-	10	11	(0)	-	-	(2)	-
	SNOW[4]	12	12	5	7	(1)	13	-	12	4	(2)	-	-	(2)	-
1817	PREC.	12	17	12	13	18	16	88	12	16	9	-	-	$15^{\underline{28}}$	-
	RAIN	1	7	12	13	15	2	50	1	11	9	-	-	0	-
	SNOW[4]	11	12	0	0	4	15	42	11	5	0	-	-	15	-
1818	PREC.	9	12	16	-	-	$12^{\underline{29}}$	-	8	12	19	12	11	12	74
	RAIN	6	12	16	-	-	9	-	6	12	19	12[1]	11	5	65
	SNOW[4]	3	1	0	-	-	3	-	2	0	0	0	0	9	11
1819	PREC.	3	15	11	9	15	21	74	6	9	12	15	12	21	75
	RAIN	2	11	11	9	14	6	53	4	8	12	15	11	7	57
	SNOW[4]	1	10	1	0	3	19	34	3	3	0	0	2	17	25
1820	PREC.	20	15	-	-	(10)	$(9)^{\underline{25}}$	-	15	11	15	18	18	12	89
	RAIN	8	14[1]	(6)	(5)	(6)	(5)	-	8	11	15	18	16	7	75
	SNOW[4]	15	2	(0)	(0)	(4)	(5)	-	10	0	0	0	6	7	23
1821	PREC.	-	-	-	-	-	-	-	8	5	(6)	-	-	-	-
	RAIN	(1)	(1)	(2)	-	-	-	-	6	4	(6)	-	-	-	-
	SNOW[4]	(1)	(4)	(1)	-	-	-	-	2	1	(0)	-	-	-	-

[1] Including hail (1 day, BR; 2 at EM).

[2] Underlined superscripts: number of days with observations, where incomplete.

[3] Bracketed numbers based on Journals only; where information fragmentary, "precipitation" omitted.

[4] Snow includes all forms of solid precipitation, including that "hail" which appears to refer to ice pellets associated with instability showers from the Bay. "Showers" noted in the Registers and Journals appear to refer to rain showers.

PRECIPITATION[5]		M	J	J	A	S	O	Total
GREAT WHALE	1941-70	14	11	15	16	18	18	92
FORT GEORGE	1941-70	7	7	10	11	12	12	59
EASTMAIN	1960-72	9	9	13	13	17	15	76
	1951-80	9	10	14	12	15	14	74

[5] Sources: Atmospheric Environment Service data tabulations, 1973, 1982.

Wind Direction

Owing to the presence of the Bay, the radiation input for a given summer along this coast is particularly sensitive to both regional and local wind direction. Although months from May to August which are colder than normal often show a higher proportion of Bay weather - onshore winds, low cloud, high humidities, fog, precipitation - abnormally low temperatures can also occur when very clear weather is associated with brisk northerly flow. These conditions foster large differences in microclimates. Typically, offshore winds during these months bring clear or partly cloudy skies, dry conditions and warmer weather.

The frequency of wind direction patterns for the 1814-1821 period were not unlike those for similar anomalously cold, warm or average months today. There was, however, an unusually high frequency of north winds at Whale River during the record cold June 1815 and in June and July 1816 (Figure 4). Normally north winds prevail at Great Whale in May and early June, associated with the series of anticyclones which cross from the Arctic into southeast or eastern Canada at that time, but this is usually superseded by prevailing westerly flow, which is characteristic of summer. Although west winds were more frequent than north in July 1816, the pattern remained akin to that of spring until the end of the month. This suggests the unseasonal persistence of high pressure over Hudson Bay. The end of July saw a change in the weather patterns at Whale River (Figure 4). In August, 58% of the winds were from the west, and no north winds were reported. Through July and August, there was little southerly flow, and the land component remained small.

At Eastmain in June 1815, May and June 1816 northerly winds retained their seasonal prominence, but winds from the south were frequent in June 1816. This was associated with two or three short spells of warmer weather; these were reflected at Whale River but were briefer. No Register was kept at Eastmain during the rest of the season, but Russell noted six days in July with very strong northwest winds, on five of which they may have reached gale force.

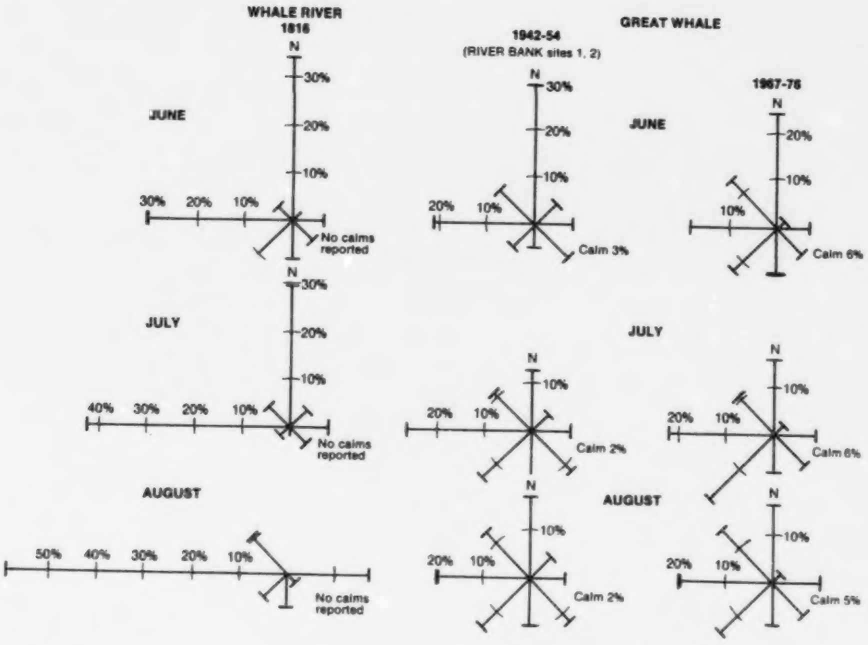


FIGURE 4: *Windroses for Whale River 1816, and Great Whale 1942-1954, 1967-1976. (The recent river bank sites are comparable with that of 1816.)*

In contrast, the pattern of wind direction at Big River, May through August 1817 (Figure 5), indicates a succession of depressions -- more frequent flow from the westerly sector in general, and the southwest in particular. While May and June retain the characteristic northerly flow, the frequency is less marked. As at Whale River in 1816, there was an unusually high occurrence of winds off the Bay from June to August 1817. In September, wind directions were more variable, with emphasis on southeast-to-south flow, but by October 1817, west winds prevailed. An equinoctial storm of rare intensity occurred on October 8, when a westerly gale, combined with an extremely high tide, washed away some 1.8 m from the river bank, flooded the lower half of Governor's Island, upset the hay stacks, and blew down trees.

From May through July 1817, Eastmain shows an increase in the number of southwest winds compared with 1816, and in the total number of winds from the Bay. It is suggested that the sites of Big River and Eastmain on opposite shores of their rivers may be responsible for a degree of channelling with respect to southwesterly and northwesterly flow, respectively.

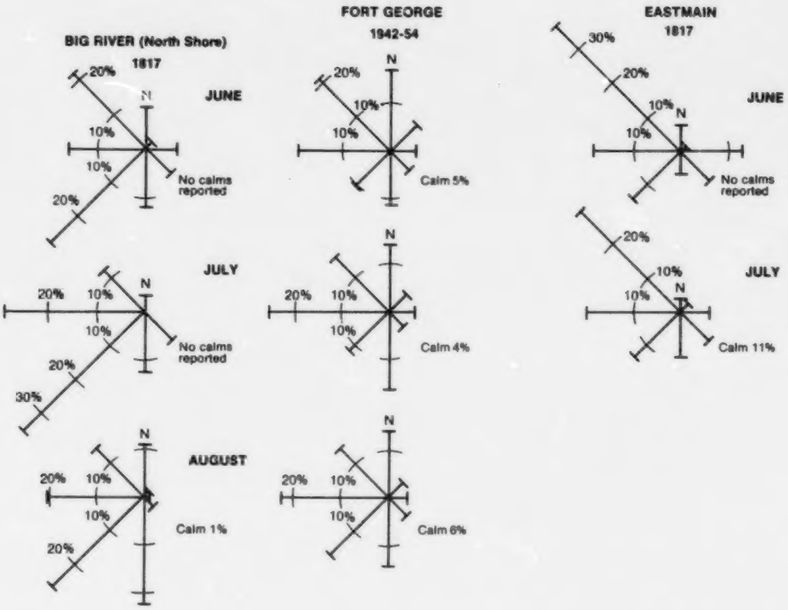


FIGURE 5: *Windroses for Big River and Eastmain 1817, and Fort George 1942–1954. (Until August 1817, Big River Post was situated on the north shore of the river; in September 1817, it was moved across the river to Governor's Island, to what is now the modern site of Fort George.)*

There is little difference in the wind distribution between the two Posts in October 1817. The major storm of October 8 was also experienced at Eastmain, where the lime kiln was washed away.

Months which are abnormally warm at Great Whale and Fort George often show both a higher frequency of winds from east southeast to south, and of land winds in general. This is seen at Big River in the warm months May 1818, July and August 1819 and June 1820. In May 1818, the prevailing winds at Big River were from the south. At Eastmain, May 1818 and July 1819, warmer than the modern average, again had a higher percentage of winds from the land (50% and 61% respectively), with some 26% of all winds from east southeast to south southwest.

Cloudiness

A striking feature of all these early nineteenth century summers is the greater amount of "clear" weather (0 to 5 tenths cloud cover), especially from May to August (Figure 6). Today, low cloud is dominant along this coast. In an average season at Great Whale similarly clear skies can only be expected at about one third of these morning, noon and evening hours from May to July, with increasing cloudiness later in the season. In exceptionally cold months, there is typically a decrease in the number of clear hours.

In 1816, the frequency of clear weather at Whale River was not only greater than the present average, but about twice what might be expected today in very cold months. A listing of clear hours against wind direction indicates a high frequency of northerly flow on these occasions, and an unusually large number of west winds in July. Over the four months, nearly two thirds of these hours occurred when the winds were from the Bay. I suggest that the clear weather in 1816 was in part the result of: (1) the prolonged influence of arctic airmasses at this period; (2) the greater frequency, persistence and intensity of spring anticyclones over Hudson Bay extending through July; and (3) probably not unrelated, the late break-up and the unusual persistence of heavy compacted ice in much of Hudson/James Bay through the summer.

The synoptic situation at Big River and Eastmain in 1817 apparently differed from that of 1816 in the more frequent passage of depressions. At Big River, the frequency of clear hours was rather lower in May and June, but remained high in July and August. The break-up and clearing of the Bay ice was again exceptionally late. There are no modern data for comparison, but it is interesting to note that although clear weather was associated with

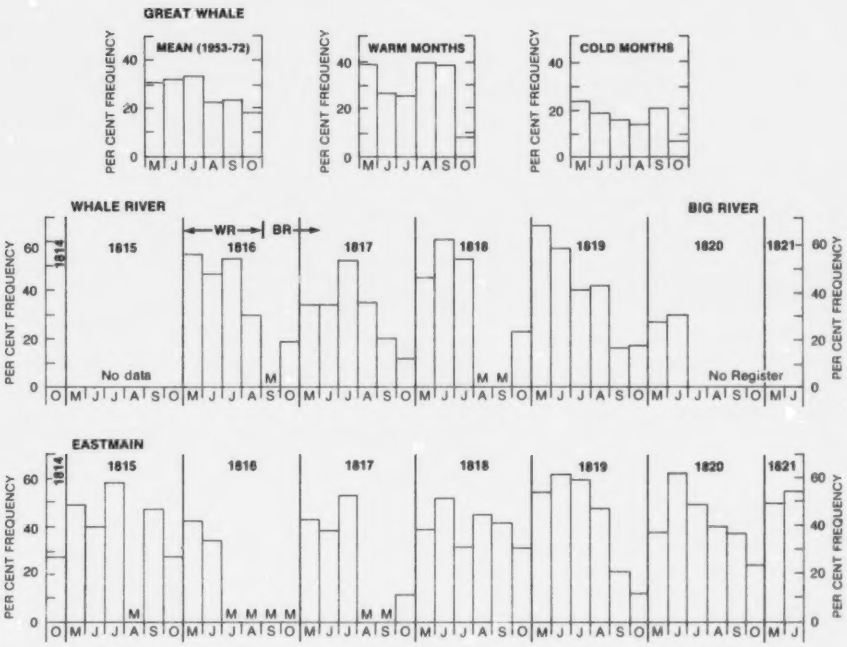


FIGURE 6: *Whale River, Big River, Eastmain, 1814-1821: relative frequency of "clear" hours, together with reference values for average, warm and cold months at Great Whale. Morning, noon and evening hours. (A "clear" hour is defined by 0 to 5 tenths cloud cover.)*

north winds in May 1817 at Big River, from June to August it tended to occur most often with winds from the southwest, and in July with all westerly components. Cloudy skies (6 to 10 tenths cloud cover) in 1816 and 1817 occurred predominantly with winds off the Bay, which is what can be expected at this season today.

DISCUSSION

Summers 1815 to 1820 - Volcanic Aerosol and Sunspots

From early this century, many have speculated on the coincidence of the widespread depression of temperature in the summer of 1816, the presence of volcanic aerosol arising from the eruptions which culminated in that of Mt. Tambora (8°S, 118°E) in April 1815, and the abnormaly low sunspot count at that period.

Empirical studies and certain theoretical considerations[1] have suggested that *in higher latitudes* any lowering of air temperature near the surface, following a massive injection of volcanic matter into the stratosphere, might be expected to be most apparent in the warm season, and of greater magnitude than in lower latitudes. The depletion of incoming solar radiation might be greater owing to the oblique angle of the beam, and a significant contributing factor at this season could be the effect on air temperature of increased or lingering sea ice and snow cover. Certain analogues concerning the circulation and residence time of stratospheric aerosol and its eventual fallout are provided by the studies of radioactive debris in the 1950s and early 1960s. It appears that the timing of the onset of any temperature loss would depend both on the latitude and the time of year of the eruption. Where the volcanic activity occurs in low latitudes as in 1815, a lag would be expected. Lamb (1970) has suggested, in keeping with the seasonal changes of the stratospheric circulation, that the poleward drift of the matter from lower latitudes proceeds with spurts each autumn. Once the material has entered the polar stratosphere, the residence time may be two or three years or more, hence any lowering of temperature in arctic/subarctic regions could last for more than one year. Recently Rampino and Self (1982) put forward the idea that the sulphate component may be more effective than silicate dust in decreasing surface temperature for several years following volcanic eruptions. They also suggest that there may be a limit to such stratospheric loading.

---

[1] See for example: Lamb 1970; Mass and Schneider 1977; Taylor *et al.* 1980; Self *et al.*1981; and, for a review of radioactive fallout in northern regions, Wilson 1967.

The eruption of El Chichon (17°N, 93°W) in April 1982 provides the first major opportunity to use modern instrumentation to monitor and measure the composition, dispersal and modification of the volcanic cloud, and the effect of the aerosol on the energy exchanges in the atmosphere and at the surface. Early results[1] have shown the gradual northward progress of the cloud through the summer of 1982, and its detection near the north polar stratosphere in September and October. With time, the cloud appeared to grow larger and denser, and interest has become increasingly focused on the sulphur component, and the possible role of high-level photochemical conversion from gas to acid particles. While measurements show an energy loss at the surface of as much as 10% to 25% through depletion of the direct solar beam, there is substantial compensation through scattering and increased diffuse radiation, reducing the net loss in solar radiation to 2% to 5%. This compensation had been noted following the lesser eruption of Mt. Agung (8° 30'S, 115° 30'E) in February/March 1963 (Bradley and England 1978). There is also satellite evidence of enhanced infrared emission from the cloud, but the debate as to the degree to which the aerosol layer may increase the infrared balance at the surface continues.

Examining the events from 1815 to 1820 along eastern Hudson/James Bay in this context, there was the onset in late autumn 1815 of consecutive winter and summer seasons of exceptional cold lasting until late winter 1818. For the critical summers 1816, 1817, a year's delay could be invoked and the extraordinary lowering of temperature occurred two years running. The degree of winter cold culminated in January/February 1818. The -6°C anomaly at Whale River in 1816 was greater than that recorded farther south in the northeastern United States and in central England, but was in keeping with the lowering of snowline at this time on north-facing slopes of the Alps (C. Pfister, personal communication). At Great Whale, during the coldest July/August on modern record (1965), the arctic/subarctic boundary lay, as in 1816 and 1817, close to Eastmain. Even then, the July temperature was 2°C warmer than in 1816 (Figure 2), and although July to October temperatures were well below normal, the two spring months were near average, and the season remained an isolated case.

---

[1] Correspondence and announcements in the *Bulletin of the American Meteorological Society*, 1982, 63:1314, 1437; 1983, 64:393-394. Also *Chinook*, Summer 1982, p. 38.

There is the matter of clearer skies in 1816, 1817 compared with the very cold summer months along this coast today. This implies a potentially larger receipt of solar radiation at the surface. Yet, the ineffectual heating of the air at "screen" level can most likely be accounted for, without recourse to direct solar depletion, by: frequent arctic airmasses; cold advection from the north or from the ice; reflection from late snow and ice; many fresh snowfalls; and heat energy directed through summer into the melt of snow and ice and the thawing and drying out of the soil. This points more generally to the complexity of weather processes along this cloudy, windward, subarctic/arctic coast: to the many transformations of energy which occur before any effect of stratospheric aerosol can be translated into a change in seasonal air temperature near the surface. The differential distribution of the aerosol by the atmospheric circulation is another consideration. It is perhaps more in keeping with our present experience of climatic variability, that the chief effects of volcanic activity might rather be indirect in middle and high latitudes, and vary from region to region.

Two further questions remain. Firstly, what changed in spring 1818? What might have brought about the spectacular return to the "normal" mode for the early nineteenth century? The sudden swing to near-record warmth in April/May 1818, following hard upon the coldest weather recorded at this time, almost suggests an "over-compensation" in redressing the balance. Had there been a significant scavenging of the aerosol by the end of the very wet 1817 season? Secondly, as was the case in New England, *unusually cold weather was in evidence along this coast before April 1815*. According to Alder (HBC, B77/e/3), and supported by entries in the Eastmain Post Journals, the region had experienced a run of very unfavourable seasons marked by periods of intense cold since about 1811/1812. In 1811, the ship *Prince of Wales* was prevented from returning to England by ice blocking Hudson Strait on October 25, and had wintered in Strutton Sound. While a number of smaller eruptions did take place in the years preceding Tambora, there is also the extraordinary concurrent event of the double cycle of abnormally low sunspot number, which spanned the first two decades of the nineteenth century (Eddy 1976, p. 1191).

A direct relationship between sunspot number and weather and climate has not yet been demonstrated, nor have mechanisms been isolated (for a review, see Pittock 1983). A study of the probable effects of variations in the ultraviolet and corpuscular radiation received from the sun has formed one avenue of research. Given the apparent connection between auroral/geomagnetic activity and sunspots, it is worth noting the location of Great Whale today with respect to the auroral oval (McCormac 1967). In comparison with the high frequency of auroral activity observed in recent years, there was no mention of an aurora along this coast in the period under study. There remains the intriguing idea put forward by Rampino *et al.* (1979) that an abrupt depression of air temperature following a major volcanic eruption, such as Tambora, often occurs when a cooling trend is already underway. They suggest that changes in atmospheric circulation caused by solar variations induce additional crustal stresses, and that a coincidence of cool climate and some large explosive eruptions may both result from a common underlying cause.

Autumn 1815 to Spring 1818, and the Inception of Permanent Snowfields over New Québec/Labrador

In the growth and expansion of continental ice sheets, there now appears to be a general consensus on the initial importance of the coalescence of areas of snowcover and drift which survive from one winter to the next. Even in modern seasons small residual patches of snow have been reported on the higher, more rugged parts of the east coast of Hudson Bay (Hare 1951) and on the New Québec/Labrador plateau (Ives 1960), and the conditions for survival seem complex. That such conditions may have been approximated in the Little Ice Age has long been postulated (for a review, see Barry *et al.* 1975). The years 1816 and 1817 came closer than any presently on record to the survival of extensive areas of snow and drift through the summer along, and in the immediate hinterland of, the east coast of Hudson/northern James Bay, and possibly on the plateau itself. These years thus provide useful observed data to set against some of the current discussion and hypotheses.

Lowering of Summer Temperature

Over the past 30 years a number of researchers have proposed a reduction in mean daily temperature of about 5° to 6°C for the formation of a perennial snowcover on the central plateau (for a summary, see Williams 1979). Williams has called for a lowering of spring and summer temperatures by as much as 10° to 12°C to achieve extensive glacierization over New Québec/Labrador, based on an energy budget model. Loewe (1971) too considered 6°C not enough unless coupled with increased precipitation. On the other hand, Manley (1949, 1975) suggested that only minor summer cooling (within present climatic variability) might produce permanent snows in maritime areas between 50° and 60°N, if such conditions were to persist over a number of decades, or more.

In midsummer 1816, the reduction in mean daily temperature at Whale River was about 5° to 6°C; at Big River in 1817 about 4.5°C, but both here and at Eastmain greater than 6°C in spring and fall. The 6°C negative anomaly in July 1816 brought the absolute mean temperature to below 5°C - just over 2°C lower than the modern record in 1965 (Figure 2, Table 1). The median height of the freezing level above Great Whale in July 1965 was some 600 to 700 m lower than the modern 10-year value (Titus 1968). This suggests that the -6°C anomaly in July 1816 may have produced a 1000 m drop (taking into consideration a higher frequency at low levels), to bring the freezing level to within 1500 m of the surface.

While recognizing the danger in extrapolating the influence of the Bay inland beyond the coastal zone, it is noteworthy that in summer and autumn 1965, the negative anomalies in the mean temperature at Schefferville and Nitchequon closely paralleled those at Great Whale, and spring was colder inland. At Nitchequon, the median height of the freezing level in July 1965 was about 500 to 600 m lower than the decadal value. In July 1816, it may have been 800 to 900 m lower.

However, such extreme reductions were only achieved for two years, 1816 and 1817. There remain critical questions such as how many consecutive years of "extreme" seasonal weather would be required to establish an effective snowcover, and whether such persistence itself is within our present experience of atmospheric variability. Manley pointed out that, at present, disturbances do not appear to be sustained for more than four or five consecutive years at most. Could this be enough?

Increased Snowfall

The lowering of temperature along the coast was combined with a substantial increase in snowfall. This was concentrated in late winter to late spring and early autumn to early winter; there was little change in mid-winter precipitation. Largest increases occurred along James Bay. While this tendency was present from 1814 to 1821, it was most marked from autumn 1815 to winter 1818. At Whale River in 1816, there were only five weeks without measurable snowfall (mid-July to mid-August). Manley (1975) and Bowling (1975) have underlined the importance of late spring and early summer snowfall in the development of perennial snowcover. Hare (1951) speculated on the influence of the late autumn/early winter snowbelt (related to the open Bay) on the formation of permanent snowfields along this east coast. With the lowering of autumn temperature in the historical period, the snowbelt was more prominent, with a southward extension of the zone of maximum snowfall from Great Whale along James Bay towards Eastmain. The development at an early stage of glaciation of a zone of perennial snowcover on the east coast of the Bay (Hare 1951) would appear quite probable.

Atmospheric Circulation and Ice

Weather events in early June 1816 have received close scrutiny in New England (e.g. Ludlam 1966), owing to the late occurrence of frost and snow accompanied by strong northwest winds. Looking at the weather "upstream" at the Posts on the Bay and inland at Neoskweskau[1], the sequence of weather appears to have been associated with the intensification of an anticyclone over the ice-covered surface of Hudson Bay (Figure 7) - a classic case of a Hudson Bay high as described by Johnson (1948). Such a build-up of surface pressure occurs under the influence of a warm upper ridge, with troughs located down over the western United States and off the east coast. The wave-length is thus shorter than normal, and the zonal index below average. The pattern can persist for up to five days or more. Johnson found this situation to be preceded by blocking over the North Atlantic or western Europe (see also Treidl *et al.* 1918). Today, the Hudson Bay high is characteristic of the early spring months, March to May. Evidence suggests that in 1816, it may have occurred with unusual frequency through June and July. If this were so, the mean ridge over west-central Canada, shown on Lamb and Johnson's (1966, p. 57) map of mean surface pressure for July 1816, should probably have its axis closer to Hudson Bay.

There can be no doubt that the late presence of large quantities of ice in the eastern half of the Bay (this intricately related to the atmospheric circulation[2]) played a largepart in the supression of summer temperature on this coast in 1816, just as the ice and low sea-surface temperature were to affect western Europe at this time (Lamb 1977; Manley 1975). Lamb and Johnson's (1966) surface pressure map for the North Atlantic region,

---

[1] Neoskweskau is about 450 km up Eastmain River.

[2] See, for example, a study published by Catchpole and Faurer (1983).

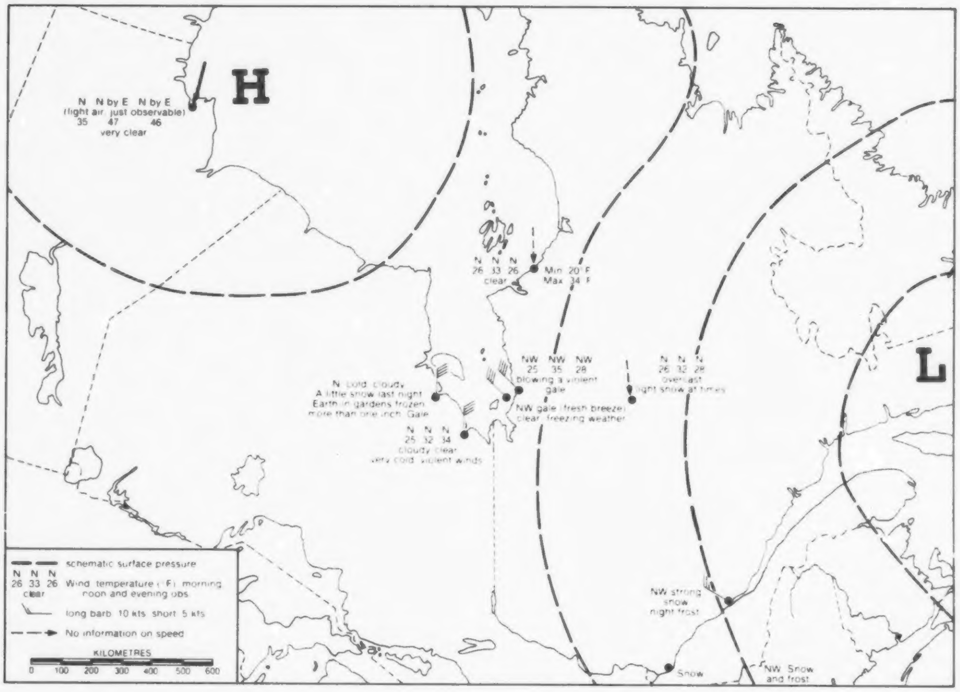

H
N N by E N by E
(light air, just observable)
35 47 46
very clear
N N N
26 33 26
clear
Min 20°F
Max 34°F
NW NW NW
25 35 28
blowing a violent gale
NW gale (fresh breeze)
clear, freezing weather
N N N
26 32 28
overcast
light snow at times
L
N, cold, cloudy
A little snow last night
Earth in gardens frozen
more than one inch. Gale
N N N
25 32 34
cloudy clear
very cold, violent winds
NW strong
snow
night frost
Snow
NW Snow
and frost
schematic surface pressure
N N N
26 33 26
clear
Wind, temperature (°F) morning, noon and evening obs
long barb 10 kts short 5 kts
No information on speed
KILOMETRES
0 100 200 300 400 500 600

FIGURE 7: *Schematic presentation of the synoptic situation on June 6, 1816.*

summer 1816, shows deep troughs off the east coast of North America and over the west coast of Europe, with the Atlantic ridge pushing north from east Greenland to Iceland. With this in mind, the Minutes of Council of the Royal Society, November 20, 1817 (Appendix 2), are of great interest. The Committee noted a considerable change in climate in the polar region during the two previous seasons, in that the severity of the cold of past centuries had abated. Evidence lay in the opening of the Greenland Sea between 74° and 80°N (the Greenland coast was now accessible by sea) and in the unusually large number of "ice islands" brought down into the Atlantic through Davis Strait during 1816[1] and 1817; there had also been an exceptional quantity of ice off the coast of northern Ireland in 1817, which had remained unthawed until the middle of August. Was the "warming" a further instance of the unusually strong meridional circulation of these two seasons? Was there an additional energy source related perhaps to submarine coastal activity in the Arctic (Rampino *et al.* 1979)?

The Social and Economic Impact of These Seasons[2]

This was a time of political, economic and social instability in England and difficult for business enterprise. The Governors of the Hudson Bay Company introduced a number of measures aimed at increasing the overall efficiency and cost-effectiveness of its operation in Canada and at weathering the storm. These included: (1) a northward extension of permanent settlement to Whale River in 1814, with a view to diversification of the trade in the direction of the whale oil business, and to make contact with the Eskimo peoples; (2) the cutting of expenditure in costly European food by growing vegetables and rearing cattle at the Posts, and putting more emphasis on local country produce (fish and game). It turned out to be a catastrophic period to experiment either with local agriculture or northward expansion!

The unrelenting sequence of cold seasons, winter and summer alike, from autumn 1815 to spring 1818, were a test of sheer survival for both white and native peoples, and a warning of the dangers in trying to live from the land in such marginal areas without adequate cushioning. During this period, the gardens failed, fish and game were exceptionally scarce, and the very late break-up and early freeze-over of the Bay and rivers threatened both the life-line with England and the internal distribution of food and trade. The situation was worsened by the wintering over of ships from England during 1815/1816 and 1816/1817, trapped by the early blocking of Hudson Strait, which increased the demands on the rapidly diminishing reserves of food. Some sailors died of scurvy, and many native people from starvation and cold. For the minimal heating of the Hudson Bay Company Posts, it required nearly twice as much firewood at Whale River from May to August 1816 than the corresponding average today; wood was not easily procured, and the men were sick and weak. In the 1817 season at Big River and Eastmain, fuel needs more than doubled. The Bay communities did not escape the distress which these years of unusual climate brought to many other peoples in the Western world and elsewhere (Post 1977). However, the quality of the furs was excellent.

---

[1] Newell (1983) indicates that in 1816 the ice off the Labrador coast (55° to 59°N) was the most severe ever reported and may have lasted through the season.

[2] A detailed account of the climatic impacts of the 1814-1821 seasons is given in Wilson (1983B).

In reading the Hudson's Bay Company Journals, it is possible to see the southward expansion of snow and ice forcing back the northern margins of habitation. Whale River was abandoned in 1816. The Indians too were living here near the northern limits of their natural environment and, in unfavourable times, many now depended for survival on emergency rations from the Posts. Had ice conditions deteriorated further, as was feared in 1817, and the annual supply ships from England failed to reach the Bay, the consequences could have been grave.

SUMMARY

The wealth of weather information in the Hudson's Bay Company archives from 1814 to 1821 provides a unique opportunity to study summer climate along the east coast of the Bay during a critical decade in the Little Ice Age. These years also include the eruption of Mt. Tambora in April 1815, and span the second of a double cycle of low sunspot number - the lowest since the Maunder Minimum.

It was a period of wide variability between the seasons. There is evidence that the summers of 1816 and 1817 were not only colder than those on modern record, but were exceptionally severe even for the period. Those for 1818 and 1820 were milder and considered very favourable, suggesting that the average expectation was lower than the present-day normal. In general, the historical seasons differed from the modern in colder springs and autumns, and more frequent spring and autumn snowfall. An unexpected feature of this period was a higher proportion of clear weather (0-5/10 cloud) than would be expected today.

From autumn 1815 through April 1818, this coast experienced arctic conditions; in the consecutive summers 1816 and 1817, the arctic boundary (after Köppen) lay close to Eastmain. The mean daily temperature at Great Whale in July 1816 was nearly 6°C below the 1941-70 normal, and at Fort George in 1817, a -5°C anomaly was sustained through the season. Bay ice in 1816 and 1817 remained longer than in any year on modern record, and 1816 provides a marginal case for the carry-over from one season to the next. With its summer snowfall, 1816 also provides a marginal case for a residual snowcover on the east side of Hudson Bay. This summer had and unusually high frequency of northerly winds; the synoptic situation over Hudson Bay and eastern Canada in early June was rather like that simulated by Williams (1975) for a July snowcover.

This was a period of political, economic and social instability in England and difficult for business enterprise. The impact of the unfavourable weather from 1815 to 1817 on these marginal Bay communities made it virtually impossible to implement some of the Company's remedial measures. It was a time of starvation and great distress among the native people.

Three aspects of the results are put forward for discussion: (1) The period autumn 1815 to spring 1818 with respect to the conditions required for the development of permanent snow and ice over New Québec/Labrador; (2) The summers 1815 to 1820 and volcanic aerosol; and (3) The social and economic impact (cf. Post 1977).

ACKNOWLEDGEMENTS

I thank the Canadian Climate Centre, Atmospheric Environment Service, for continued funding for this study, and the many individuals there who have given me unfailing help and support in different ways. I feel indebted to the Hudson's Bay Company Officers at Whale River, Big River and Eastmain Posts, and others, who maintained regular weather observations through times of great distress and hardship. My appreciation and thanks go to Brian Taylor who drafted Figures 2 to 7 and to Mrs. Valerie Moore for typing the manuscript. Figure 1, drafted by Edward Hearn, has been reproduced from *Syllogeus* No. 49, p. 145.

REFERENCES

Barry, R.G., J.T. Andrews, and M.A. Mahaffy. 1975. Continental ice sheets: conditions for growth. Science 190:979-981.

Blasing, T.J., and H.C. Fritts. 1975. Past climate of Alaska and northwestern Canada as reconstructed from tree rings. *In*: Climate of the Arctic. Edited by: G. Weller and S.A. Bowling. University of Alaska, Geophysical Institute. pp. 48-58.

Bowling, S.A. 1975. Possible significance of recent weather and circulation anomalies in northeastern Canada for the initiation of continental glaciation. *In*: Climate of the Arctic. Edited by: G. Weller and S.A. Bowling. University of Alaska, Geophysical Institute. pp. 91-97.

Bradley, R.S., and J. England. 1978. Recent climatic fluctuations of the Canadian High Arctic and their significance for glaciology. Arctic and Alpine Research 10:715-731.

Canada, Atmospheric Environment Service. 1973. Temperature and precipitation, 1941-1970, Québec. 65 pp.

______. Canadian climate normals, 1951-80, temperature and precipitation, Québec. 216 pp.

Catchpole, A.J.W., D.W. Moodie, and D. Milton. 1976. Freeze-up and break-up of estuaries on Hudson Bay in the eighteenth and nineteenth centuries. Canadian Geographer 20:279-297.

Catchpole, A.J.W., and T.F. Ball. 1981. Analysis of historical evidence of climatic change in western and northern Canada. *In*: Climatic Change in Canada 2. Edited by: C.R. Harington. Syllogeus No. 33:48-96.

Catchpole, A.J.W., and M.A. Faurer. 1983. Summer sea ice severity in Hudson Strait, 1751-1870. Climatic Change 5:115-139.

Cooke, A., and C. Holland. 1978. The exploration of northern Canada, 500 to 1920, a chronology. The Arctic History Press, Toronto. 549 pp.

Cropper, J.P., and H.C. Fritts. 1981. Tree-ring width chronologies from the North American Arctic. Arctic and Alpine Research 13:245-260.

Eddy, J.A. 1976. The Maunder minimum. Science 192:1189-1202.

Hare, F.K. 1951. The present-day snowfall of Labrador-Ungava. American Journal of Science 249:654-670.

Hudson's Bay Company (HBC). References to the HBC archive catalogue.

Ives, J.D. 1960. Semi-permanent snow beds in central Labrador-Ungava. *In*: Field Research in Labrador-Ungava. McGill University, Montréal, McGill Subarctic Research Papers No. 9:45-55.

Johnson, C.B. 1948. Anticyclogenesis in eastern Canada during spring. Bulletin of the American Meteorological Society 29:47-55.

Lamb, H.H. 1970. Volcanic dust in the atmosphere. Philosophical Transactions, Royal Society, London, Series A, 266:425-533.

______. 1977. Climate, present, past and future. Volume 2. Climatic history and the future. Methuen, London. 835 pp.

Lamb, H.H., and A.I. Johnson. 1966. Secular variations of the atmospheric circulation since 1750. Meteorological Office (London), Geophysical Memoirs No. 110:1-125.

Loewe, F. 1971. Considerations of the origin of the Quaternary ice sheet of North America. Arctic and Alpine Research 3:331-344.

Ludlam, D.M. 1966. Early American winters 1604-1820. American Meteorological Society, Boston. pp. 190-194.

Manley, G. 1949. The snowline in Britian. Geografiska Annaler 31:135-153.

______. 1975. Fluctuations of snowfall and persistence of snowcover in marginal - oceanic climates. Geneva. World Meteorological Organization. In: Proceedings of the WMO/IAMAP symposium on Long-term Climatic Fluctuations. Norwich, August 18-23, 1975. pp. 183-188.

Mass, C., and S.H. Schneider. 1977. Statistical evidence on the influence of sunspots and volcanic dust on long-term temperature trends. Journal of Atmospheric Science 34: 1995-2004.

McCormac, B.M. 1967. Aurora and airglow. Reinhold, New York. 689 pp.

Newell, J. 1983. Preliminary analysis of sea-ice conditions in the Labrador Sea during the nineteenth century. In: Climatic Change in Canada 3. Edited by: C. R. Harington. Syllogeus No. 49:108-129.

Parker, M.L., L.A. Jozsa, S.G. Johnson, and P.A. Bramhall. 1981. Dendrochronological studies on the coasts of James Bay and Hudson Bay. In: Climatic Change in Canada 2. Edited by: C.R. Harington. Syllogeus No. 33:129-188.

Payette, S. 1976. Succession écologique des forêts d'épinette blanche, et fluctuations climatiques, Poste-de-la-Baleine, Nouveau-Québec. Canadian Journal of Botany 54: 1394-1402.

Pittock, A.B. 1983. Solar variability, weather and climate: an update. Quarterly Journal of the Royal Meteorological Society 109:23-55.

Post, J.D. 1977. The last great subsistence crisis in the western world. John Hopkins University Press, Baltimore. 240 pp.

Rampino, M.R., S. Self. 1982. Historic eruptions of Tambora (1815), Krakatau (1883) and Agung (1963), their stratospheric aerosols, and climatic impact. Quaternary Research 18:127-143.

Rampino, M.R., S. Self, and R.W. Fairbridge. 1979. Can rapid climatic change cause volcanic eruptions? Science 206:826-829.

Royal Society, London. Reference to the RS archive catalogue.

Self, S., M.R. Rampino, and J.J. Barbera. 1981. The possible effects of the large 19th and 20th century volcanic eruptions on zonal and hemispheric surface temperature. Journal of Volcanology and Geothermal Research 11:41-60.

Sowden, W.J., and F.E. Geddes. 1980. Weekly median and extreme ice edges for eastern Canadian seaboard and Hudson Bay. Atmospheric Environment Service. Ice Climatology and Applications Division, Ottawa. 4 pp. and charts.

Taylor, B.L., T. Gal-Chen, and S.H. Schneider. 1980. Volcanic eruptions and long-term temperature records: an empirical search for cause and effect. Quarterly Journal of the Royal Meteorological Society 106:175-199.

Titus, R.L. 1968. Freezing-level statistics for Canada. Canada Department of Transport, Meteorological Branch, Climatological Studies No. 12:1-40.

Treidl, R.A., E.C. Birch, and P. Sajecki. 1981. Blocking action in the northern hemisphere: a climatological study. Atmosphere - Ocean 19:1-23.

Williams, J. 1975. The influence of snowcover on the atmospheric circulation and its role in climatic change: an analysis based on results from the NCAR global circulation model. Journal of Applied Meteorology 14:137-152.

Williams, L.D. 1979. An energy balance model of potential glacierization of northern Canada. Arctic and Alpine Research 11:443-456.

Wilson, C. 1967. Radioactive fallout in northern regions. United States Army, Cold Regions Research and Engineering Laboratory (CRREL), Hanover. Cold Regions Science and Engineering Report 1-A2:1-35.

______. 1982. The summer season along the east coast of Hudson Bay during the nineteenth century. Part I. General introduction; climatic controls; calibration of the instrumental temperature data, 1814 to 1821. Canadian Climate Centre Report No. 82-4:1-223.

______. 1983 A. Some aspects of the calibration of early Canadian temperature records in the Hudson's Bay Company Archives: a case study for the summer season, eastern Hudson/James Bay, 1814 to 1821. In: Climatic Change in Canada 3. Edited by: C.R. Harington. Syllogeus No. 49:144-202.

______. 1983 B. Part II. The Little Ice Age on eastern Hudson Bay; summers at Great Whale, Fort George, Eastmain, 1814-1821. Canadian Climate Centre Report No. 83-9: 1-145.

APPENDIX 1: *A NOTE ON BAY ICE IN RECENT EXTREME SEASONS.*

Of recent years,, 1969 provides the closest analogy to the 1816-1817 seasons, with the prolonged compacting of the ice in the east and south, as a result of persistent northerly winds through spring and much of the summer. The final traces of ice did not melt completely in the southern part of Hudson Bay until September 20 - the latest on record. The latest disappearance of the ice in James Bay was recorded in the first week of September 1965. Unlike 1969, the autumn freeze-over in 1965 was advanced, with old and new ice across the exit into Hudson Strait by early October. The case was similar in 1972. During the cold autumns of 1972 and 1965, sheets of new winter ice were observed from early November over the western half of James Bay, and in 1974 in the vicinity of the Belcher Islands. The May-to-October seasons 1965, 1969, 1972 are three of the coldest registered at Great Whale; the midsummer 1965 and autumn 1974 are the coldest of the 1926-76 series.

Sources:

Canada, Atmospheric Environment Service. Ice charts for 1972, 1974, Hudson Bay and approaches. Ice Climatology and User Applications Division, Ottawa.

Canada, Meteorological Branch (Department of Transport). 1967. Ice, summary and analysis, 1965. Hudson Bay and approaches. 51 pp.

Canadian Meteorological Service (Environment Canada). 1971. Ice, summary and analysis, 1969. Hudson Bay and approaches. 47 pp.

APPENDIX 2: *ROYAL SOCIETY, LONDON. MINUTES OF COUNCIL, NOVEMBER 20, 1817.*

The President proposed a letter be written to the Admiralty recommending a voyage be undertaken to investigate the North Polar regions. The following is an extract from the draft of the letter.

> "It will without doubt have come to your Lordship's knowledge that a considerable change of climate inexplicable at present to us must have taken place in the Circumpolar Regions, by which the severity of the cold that has for centuries past inclosed the seas in the high northern latitudes in an impenetrable barrier of ice has been during the last two years greatly abated.
>
> Mr. Scoresby, a very intelligent young man who commands a whaling vessel from Whitby observed last year that 2000 square leagues of ice with which the Greenland Seas between the latitudes of 74° and 80°N have been hitherto covered, has in the last two years entirely disappeared.
>
> The same person who has never been before able to penetrate to the westward of the Meridian of Greenwich in these latitudes was this year able to proceed to 10°, 30'W where he saw the coast of East Greenland and entertained no doubt of being able to reach the land had not his duty to his employers made it necessary for him to abandon the undertaking.
>
> This, with information of a similar nature derived from other sources; the unusual abundance of ice islands that have during the last two summers been brought by currents from Davies Streights into the Atlantic. The ice which has this year surrounded the northern coast of Ireland in unusual quantity and remained there unthawed till the middle of August, with the floods which have during the whole summer inundated all those parts of Germany where rivers have their sources in snowy mountains, afford ample proof that new sources of warmth have been opened and give us leave to hope that the Arctic Seas may at this time be more accessible than they have been for centuries past, and that discoveries may now be made in them not only interesting to the advancement of science but also to the future intercourse of mankind and the commerce of distant nations".

Source:

Royal Society, London. 1817. Minutes of Council, volume 8. pp. 149-153.

APPENDED TABLES

TABLE 5: *WINTER 1814-1821: WHALE RIVER, BIG RIVER, EASTMAIN. MONTHLY MEAN TEMPERATURES (°C). WHERE UNDERLINED, RELATIVELY WARM. NOTE: THESE TEMPERATURES HAVE NOT BEEN CALIBRATED FOR DIRECT COMPARISON WITH THE MODERN DATA; HOWEVER, THEY SHOW THE RELATIVE DIFFERENCES WITHIN THE PERIOD.*[1]

MEAN DAILY TEMPERATURE		N	D	J	F	M	A	SEASON
Whale River	1815/1816	-10.7	-17.2	-25.9	-25.1	-24.3	-6.7	-18.3
Big River	1816/1817	-7.0	-23.3	-27.3	-26.7	-15.6	-9.1	-18.2
	1817/1818	-7.4	-18.9	-29.8	-28.7	-16.5	-1.4	-17.1
	1818/1819	-2.5	-18.2	-24.9	-18.3	-23.4	-4.7	-15.3
Great Whale	1941-1970	-4.7	-14.8	-22.3	-22.2	-16.5	-6.8	-14.6
Fort George	1941-1970	-4.3	-14.6	-22.6	-21.2	-14.9	-5.4	-13.8
Eastmain	1951-1980	-4.1	-15.5	-20.4	-19.7	-12.9	-2.7	-12.6

MEAN (8 AM + 5 PM T)/2		N	D	J	F	M	A	SEASON
Whale River	1814/1815	-6.1	-16.7	-28.3	-24.2	-	-	-
	1815/1816	-11.1	-18.1	-26.6	-26.4	-26.9	-8.7	-19.6
Big River	1816/1817	-8.0	-24.3	-28.8	-28.4	-17.9	-11.3	-19.8
	1817/1818	-8.1	-19.9	-30.8	-30.6	-17.6	-2.3	-18.2
	1818/1819	-3.1	-19.1	-26.3	-19.6	-25.2	-5.2	-16.4
	1819/1820	-6.7	-13.9	-24.3	-23.6	-21.9	-6.6	-16.2
Great Whale	1957-1972	-4.9	-15.9	-21.6	-22.2	-16.3	-5.9	-14.5

MEAN (6 AM + 6 PM T)/2		N	D	J	F	M	A	SEASON
Eastmain	[2]1814/1815	-5.5	-16.7	-23.3	-21.8	-18.5	-13.3	-16.5[2]
	[3]1815/1816	-10.1	-18.0	-23.8	-24.8	-22.9	-7.6	-17.9[3]
	1816/1817	-7.7	-19.1	-26.6	-26.1	-15.1	-8.9	-17.3
	1817/1818	-6.3	-19.0	-26.9	-28.3	-14.1	-2.9	-16.3
	1818/1819	-2.9	-19.1	-23.2	-16.3	-23.0	-5.1	-14.9
	1819/1820	-6.9	-15.9	-23.7	-19.6	-19.4	-5.4	-15.2
	1820/1821	-5.2	-16.8	-22.4	-23.6	-20.3	-9.2	-16.3
Great Whale	1957-1972	-4.9	-16.0	-21.7	-22.5	-17.1	-7.0	-14.9

---

[1] Sources: Atmospheric Environment Service data tabulations, 1973, 1982.

[2] Mean (8 am + 8 pm T)/2.

[3] Mean (sunrise + sunset T)/2.

TABLE 6: *WINTER 1814-1821: WHALE RIVER, BIG RIVER, EASTMAIN. NUMBER OF DAYS WITH REPORTED RAIN OR SNOW. BELOW: THE NORMAL NUMBER OF DAYS FOR GREAT WHALE, FORT GEORGE AND EASTMAIN.*

		WHALE RIVER / BIG RIVER							EASTMAIN						
		N	D	J	F	M[1]	A[2]	Total[1]	N	D	J	F	M	A	Total
1814/15	RAIN	3	–	–	–	–	M	(3)	1	–	–	–	–	–	1
	SNOW	9	8	4	6	(6)	M	(33)	10	23	11	9	9	10	72
1815/16	RAIN	–	–	–	–	–	2	2	2	–	–	–	–	6	8
	SNOW	23	12	7	8	6	8	64	20	11	5	10	11	9	66
1816/17	RAIN	2	–	–	–	–	1	3	1	–	–	–	1	2	4
	SNOW	17	9	10	3	9	6	54	14	15	7	8	8	12	64
1817/18	RAIN	5	–	–	–	1	1	7	3	–	–	–	1	3	7
	SNOW	12	13	3	7	4	9	48	10	7	7	2	2	7	35
1818/19	RAIN	3	–	–	–	–	–	3	4	–	1	2	–	1	8
	SNOW	16	14	8	6	7	7	58	16	14	7	7	8	7	59
1819/20	RAIN	3	1	–	1	1	3	9	1	–	–	1	1	3	6
	SNOW	22	8	10	8	7	7	62	15	8	9	8	9	4	53
1820/21	RAIN	–	–	–	–	–	–	–	1	–	–	–	–	3	4
	SNOW	–	–	–	–	–	–	–	12	8	9	6	8	4	47

[1] (6) based on 11 days only; other bracketed values indicate data are incomplete.

[2] M indicates data are missing.

NB. In some months where Journal remarks were not available to supplement the two or three observations at fixed hours, the counts may be conservative (cf. Manley 1978).

		GREAT WHALE[3] / FORT GEORGE[3]							EASTMAIN[3]						
		N	D	J[4]	F[4]	M	A	Total	N	D	J	F	M	A	Total
1941-70	RAIN	3	1	*	*	1	2	7	–	–	–	–	–	–	–
	SNOW	17	16	12	10	11	11	77	–	–	–	–	–	–	–
1941-70	RAIN	2	1	*	*	1	2	6	–	–	–	–	–	–	–
	SNOW	12	12	7	6	7	4	48	–	–	–	–	–	–	–
1951-80	RAIN	–	–	–	–	–	–	–	4	–	–	–	1	3	8
	SNOW	–	–	–	–	–	–	–	13	14	11	7	7	3	55

[3] Sources: Atmospheric Environment Service data tabulations, 1973, 1982.

[4] Asterisks indicate less than one day.

DAILY WEATHER MAPS FOR CANADA, SUMMERS 1816 to 1818 - A PILOT STUDY

Cynthia Wilson[1]

INTRODUCTION

A previous study[2] of the weather and climate along eastern Hudson/James Bay during the second decade of the nineteenth century (Wilson 1983B) indicated that an exceptionally cold period lasting from autumn 1815 to late winter 1818 (including the record cold summers of 1816 and 1817), had terminated abruptly in early spring 1818. At that time, monthly mean temperatures in April and May, together with some daily maxima, had approached record high values, following which the seasons had returned to near average expectation for the period. Questions immediately arise both as to the nature of the atmospheric circulation over central Canada in 1816 and 1817, and of the changes which occurred in 1818. Wind directions observed along this coast suggest that the two cold summers were themselves marked by very different frequencies of regional circulation patterns.

Owing to the particular interest of these three seasons, they have been chosen as reference periods in the present study of the feasibility of obtaining synoptic weather charts for Canada in the early nineteenth century. The Hudson's Bay Company (HBC), through its network of Posts and lines of communication in central, western and northern Canada, provided a system of synoptic weather observation unique at this time, both in the discipline imposed, the consistency of purpose and of manner of observing and recording, and in the extent of its coverage. These records form the principal data source[3].

---

[1] P.O. Box 887, Station B, Ottawa, Ontario K1P 5P9

[2] Under contract to the Canadian Atmospheric Environment Service.

[3] Weather information was abstracted with the kind permission of the Hudson's Bay Company.

CORRECTIONS:

Page 203. Last paragraph. Line 1. Should read "... was given careful consideration by Lamb ..."

NOTE: This particular aspect of the synoptic study of the Armada period was credited in error to C. Loader. The careful synoptic analysis for the months May to October 1588, published in Douglas, Lamb and Loader 1978, and Douglas and Lamb 1979, was of course the work of Professor Lamb, to whom I wish to express my apologies. (Cynthia Wilson)

Page 218. Add:

Voorhis, E. 1930. Historic forts and trading posts of the French régime and of the English fur trading companies. Department of the Interior, Ottawa. 188 pp.

THE CONSTRUCTION OF HISTORICAL WEATHER MAPS: DATA REQUIREMENTS

The modern surface weather map is essentially one of atmospheric pressure, and Kington's (1975, 1980) classic reconstruction of daily synoptic weather maps for western Europe and the northeast Atlantic in the 1780s was firmly based on a network of barometer records. The number of locations in Canada with barometer readings for the early nineteenth century appears to be very small. In the absence of pressure data, surface winds - more readily and frequently recorded - can be used to construct schematic pressure patterns (Douglas *et al.* 1978, 1979). This is not without its dangers. Regional wind direction and speed may have been modified at some sites by local topography, the geometry of the forest and other obstructions, or masked by the influence of local sea or lake breezes and valley winds. However, marked local effects can usually be detected, particularly where the network is reasonably dense. To demarcate the frontal zones separating the different air masses, the weather information should be adequate to position significant changes in winds and temperature, together with such indicators of atmospheric stability as cloud, precipitation, and thunderstorms.

The synoptic scale, that of the low-level migratory high and low pressure systems, involves disturbances of wavelength from 1000 to 2500 km. To obtain synoptic maps for the Hudson Bay region thus requires a much larger study area, with as much information upstream and downstream as can be procured. The timing and frequency of the observations are seminal to the analysis. For example, a "daily" wind direction may refer either to the morning or evening; in 12 hours, a depression moving at a rate of 50 km/hr will travel 500 to 600 km. Again, if the network is sufficiently dense, such incongruities may stand out, or a pattern of error appear with time. To ensure synchronization, modern synoptic data refer to Universal Coordinated Time (Greenwhich Mean Time). For the early weather maps, if the Hudson Bay area is considered the reference time zone, the western and eastern extremes of the map will be within ± 2 hours, which can be born in mind. In drawing up weather maps[1], "historical continuity" is a fundamental principle. The movement and transformation of airmasses, fronts, highs and lows must show a logical development from the situation on the previous chart. For the early weather maps, this becomes a prime tool in grappling with the many difficulties, including that of sparse data coverage. In order to keep track of such changes today, four sea-level charts are analyzed daily. To retain continuity in the earlier period, it would seem important to have a frequency of observations so as to permit two charts to be sketched in for each day (preferably morning and evening). At their best, the early records generally accommodate this, and often provide intermediate guidelines for midday.

---

[1] For practical information on weather map analysis, see Petterssen (1941, 1956, 1958) and Saucier (1955).

HUDSON'S BAY COMPANY WEATHER RECORDS: SUMMERS 1816 to 1818

For the trial period May to August 1816, 1817, 1818, all the Hudson's Bay Company Post Journals were searched, and direct and proxy weather information abstracted, verbatim. Further data were obtained from the Ships' Logs.

Meteorological Registers and Ships' Logs

These provide prime data. In the Post Registers, observations of air temperature, wind, weather (cloud, precipitation and extreme events), and in rare cases atmospheric pressure were recorded usually for two or three "fixed" hours a day (morning, midday, evening). In the Ships' Logs, wind and weather reports were entered at every second hour while at sea, covering the entire 24-hour period.

Wind and Weather Remarks in Post Journals and in Ships' Logs for Vessels at Anchor

The Journals and Logs often provide a brief but regular daily summary. This may include such information as to the nature of any significant changes during the day, and of the kind, intensity and timing of precipitation. Newsworthy extreme events such as frosts, gales, thunderstorms were usually noted. Some material can also be found in the logs of canoe journeys, such as the annual trips between inland and Bay Posts to exchange furs for supplies, and those undertaken to explore the hinterland. Not all Journals provide direct information on the weather, and many that do were interrupted or only cursorily written in the summer, if the Master and Clerk were absent for the trade. An advantage of the synoptic approach is that fragmented series remain very useful.

Proxy Data in the Post Journals and Logs

Items such as: (1) storm damage; (2) the prevention of different outdoor activities or travel owing to winds or precipitation; (3) remarks on local ice movement and decay or growth; (4) comments on snow cover, plant damage or growth; (5) discomfort or sickness owing to weather, indicate the nature or magnitude or timing of certain weather phenomena. Even where Journals do not contain regular weather information, such valuable items may occasionally be found.

THE QUESTION OF CALIBRATION

Space and time smoothing of some of the irregularities in the data are implicit in the map analysis.

With respect to the *temperature* data, the analysis is primarily concerned with relative differences and changes within the temperature field, rather than with absolute values. The results of an earlier study (Wilson 1982, 1983A) suggest that although the observing practices were different in the early nineteenth century, they were consistently so throughout the Hudson's Bay Company network. Thermometers were placed outside, away from the sun, about 1.5 m above the ground, and most likely on a north wall. At the morning and evening observations, and in the range near 0°C, differences between the early and modern values were probably very small. Any significant errors would be expected to stand out.

For the months under study, *barometer* readings were recorded at Edmonton in 1816 (May 2 to 9 only), Albany in 1817 and Eastmain in 1818. Although the instrument at Albany was apparently reading too low, the pressure trace remains invaluable. These data refer to station pressure; for Edmonton (altitude about 760 m) reduction to sea-level pressure is required, using the current procedure of the Atmospheric Environment Service.

Except for calms and gales, records of *wind force* were mainly confined to coastal locations at this time. At Fort Churchill and York Factory, a scale was used to indicate the force. This appears to have been the descriptive 10-point scale related to boat operations, given in the Meteorological Journal for Churchill, 1811-1813 (HBC, B42/a/139a). A probably similar scale, is also found in the Edmonton Register. It is not unlike the 13-point Beaufort scale, which although first devised in 1806 was not in general use at this time. In a number of Post Journals, and Logs, descriptive terms such as "fresh" or "strong breeze", "near" or "strong gale" appear. To convert all these indicators into an approximate speed, the Beaufort scale has been used (Atmospheric Environment Service 1970). With *wind direction*, it has been assumed that directions were recorded with respect to true north.

Other information concerning the interpretation of these wind and weather data can be found in an earlier report (Wilson 1983B). For a discussion of the use of Ships' Logs in synoptic analysis, see Oliver and Kington (1970).

DATA COVERAGE

Following Company directives sent out from England in 1814 (HBC, A6/18), a high priority was now to be given to the careful observing and recording of the weather and its impact during the active season, with a view to increasing local food production at the Posts. There were precise instructions as to the kinds of information to be noted in the Journals, and where possible a "Register of the Thermometer" was to be kept. Instruments were to be distributed to such Officers as were likely to make the observations correctly. A glance through the Hudson's Bay Company Archives for 1814/1815 shows that the program was well underway. From May to October 1815, Meteorological Registers were kept at some 11 Posts through most of the season, from Edmonton to Neoskweskau (New Québec) and Whale River to Kenogamissi (Northern Ontario).

Locations of all Posts for which Journals exist during at least part of the period May to August, 1816 to 1818, are shown on Figure 1. Data cover (both extent and type) has been assessed for each month. Table 1 offers a monthly breakdown of the number of Post Journals and Meteorological Registers, while Figures 2A-D indicate changes in spatial coverage.

The thrust of the Company's observing program carried through from 1815 into May 1816, the month which offers the best weather network for these three summers (Figure 2A). Of particular importance is the distribution of the Registers: five at Posts around the Bay: three for the interior north of the Great Lakes: one inland, to the east and two out west. (Unfortunately that at Brandon House was discontinued in March 1816.) In addition to

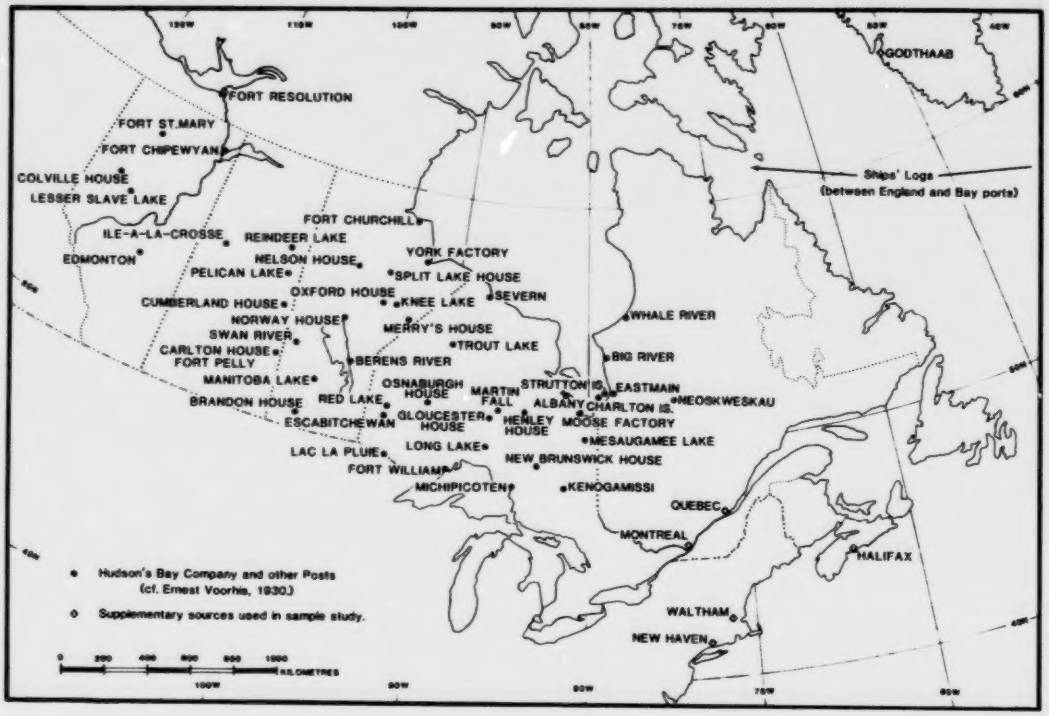

FORT RESOLUTION
FORT ST.MARY
FORT CHIPEWYAN
COLVILLE HOUSE
LESSER SLAVE LAKE
ILE-A-LA-CROSSE
REINDEER LAKE
EDMONTON
NELSON HOUSE
PELICAN LAKE
FORT CHURCHILL
YORK FACTORY
SPLIT LAKE HOUSE
OXFORD HOUSE
KNEE LAKE
SEVERN
CUMBERLAND HOUSE
NORWAY HOUSE
MERRY'S HOUSE
SWAN RIVER
TROUT LAKE
CARLTON HOUSE
FORT PELLY
BERENS RIVER
MANITOBA LAKE
OSNABURGH HOUSE
MARTIN FALL
STRUTTON IS.
EASTMAIN
BRANDON HOUSE
RED LAKE
ALBANY
CHARLTON IS.
NEOSKWESKAU
ESCABITCHEWAN
GLOUCESTER HOUSE
HENLEY HOUSE
MOOSE FACTORY
MESAUGAMEE LAKE
LAC LA PLUIE
LONG LAKE
NEW BRUNSWICK HOUSE
FORT WILLIAM
MICHIPICOTEN
KENOGAMISSI
WHALE RIVER
BIG RIVER
GODTHAAB
Ships' Logs
(between England and Bay ports)
QUEBEC
MONTREAL
HALIFAX
WALTHAM
NEW HAVEN
● Hudson's Bay Company and other Posts
(cf. Ernest Voorhis, 1930.)
○ Supplementary sources used in sample study.
0
200
400
600
800
1000
KILOMETRES

FIGURE 1: *Location of Hudson's Bay Company Posts for which Journals exist during at least part of the period 1816 to 1818. (The Figure also shows five other sites used in the sample study, June 1 to 17, 1816.)*

the 25 Post Journals, the Logs of the two ships "wintering" (until August) at Strutton Islands also contain daily weather summaries. But in June 1816, the number of Journals and Registers falls away dramatically to reach a minimum in July/August (Figure 2B): not only had the number of Posts reporting halved, but Registers had almost stopped. Above all, there was little information of any kind west of Osnaburgh House. The disruption arose from the catastrophic events associated with the battle between the Hudson's Bay Company and the North West Company for the Athabaska trade. On June 1 1816, Brandon House was plundered; the struggle moved west and continued through 1817.

TABLE 1: *TOTAL NUMBER OF POST JOURNALS; NUMBER OF METEOROLOGICAL REGISTERS.*

YEARS	DOCUMENTS	MONTHS			
1816		May	June	July	August
	Journals[1]	25	16(+3)	11(+1)	12(+2)
	Registers[2]	10(+1)	8	2(+1)	1(+1)
1817					
	Journals	12	14(+2)	12(+4)	11(+5)
	Registers	5	5	5	2(+3)
1818					
	Journals	23	16(+3)	15(+5)	13(+8)
	Registers	6	5(+1)	4(+2)	4(+2)

Through the summer of 1817, the network remained similar to that of midsummer 1816, with the addition of Berens River in June and July (Figure 2C). There were now five Registers, but confined to Bay Posts until late August, when that for New Brunswick was reactivated. The Logs of two ships wintering over at Charlton Island provide additional information for James Bay. While the 1818 season saw some recovery, notably the inclusion of Journals for Cumberland House and Carlton House, the number of Registers was not to regain the earlier level; apart from New Brunswick, these latter referred to Bay sites (Figure 2D).

Given the scarcity of material for sites in western Canada, the information recorded on canoe journeys is particularly useful along western waterways, where there was considerable

---

[1] (+3), other Post Journals containing reports from canoe trips only.

[2] (+1), additional Registers covering only a part of the month.

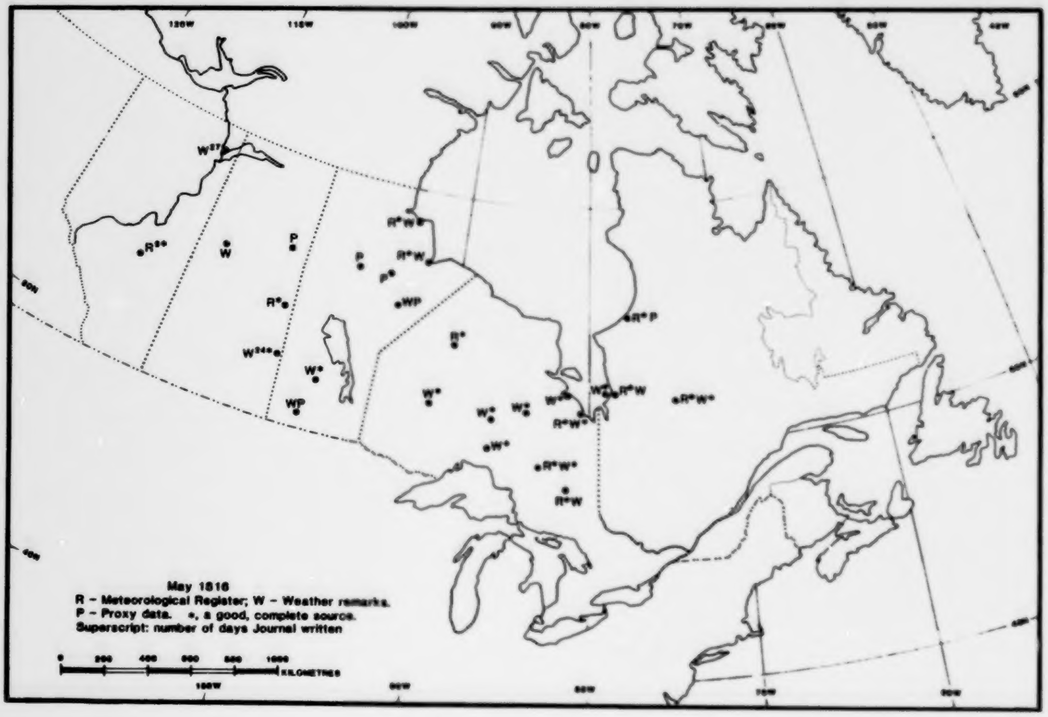

May 1816
R – Meteorological Register; W – Weather remarks.
P – Proxy data. •, a good, complete source.
Superscript: number of days Journal written
0
200
400
600
800
1000
KILOMETRES

FIGURE 2A: *Hudson's Bay Company records, weather data coverage for May 1816.*

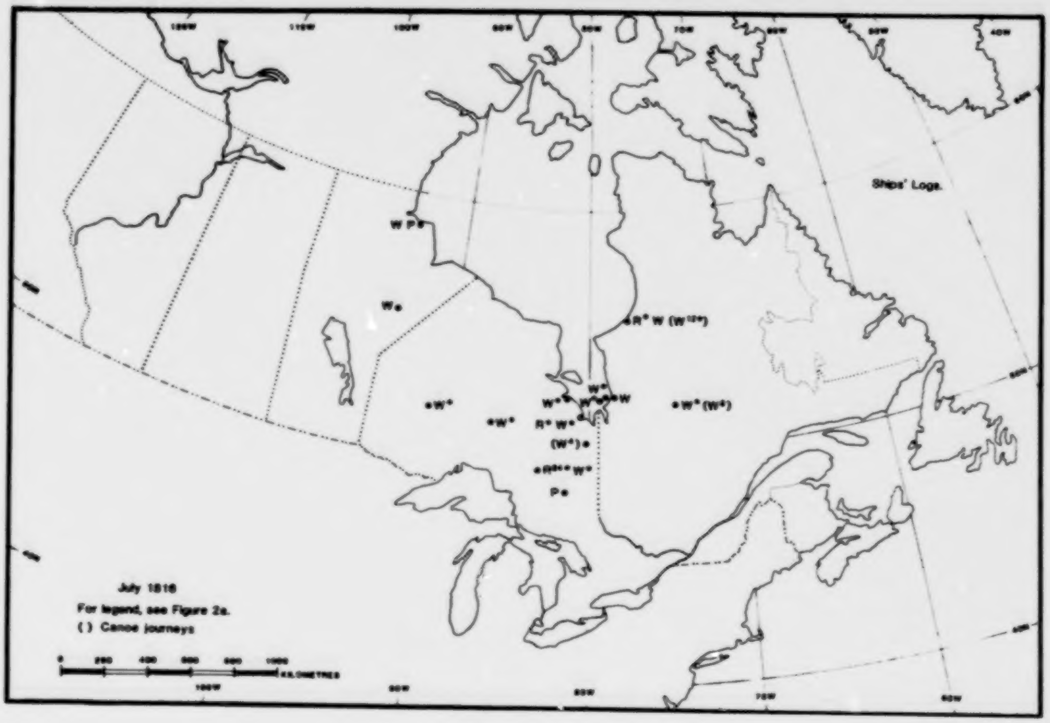

120W
110W
100W
90W
80W
70W
60W
50W
40W
Ships' Logs.
W P
W
R* W (W12*)
W*
W*
W*
W
W* (W*)
W*
W*
R* W*
(W*)
R** W*
P
July 1816
For legend, see Figure 2a.
( ) Canoe journeys
0
200
400
600
800
1000
KILOMETRES
100W
90W
80W
70W
60W

FIGURE 2B: *Hudson's Bay Company records, weather data coverage for July 1816.*

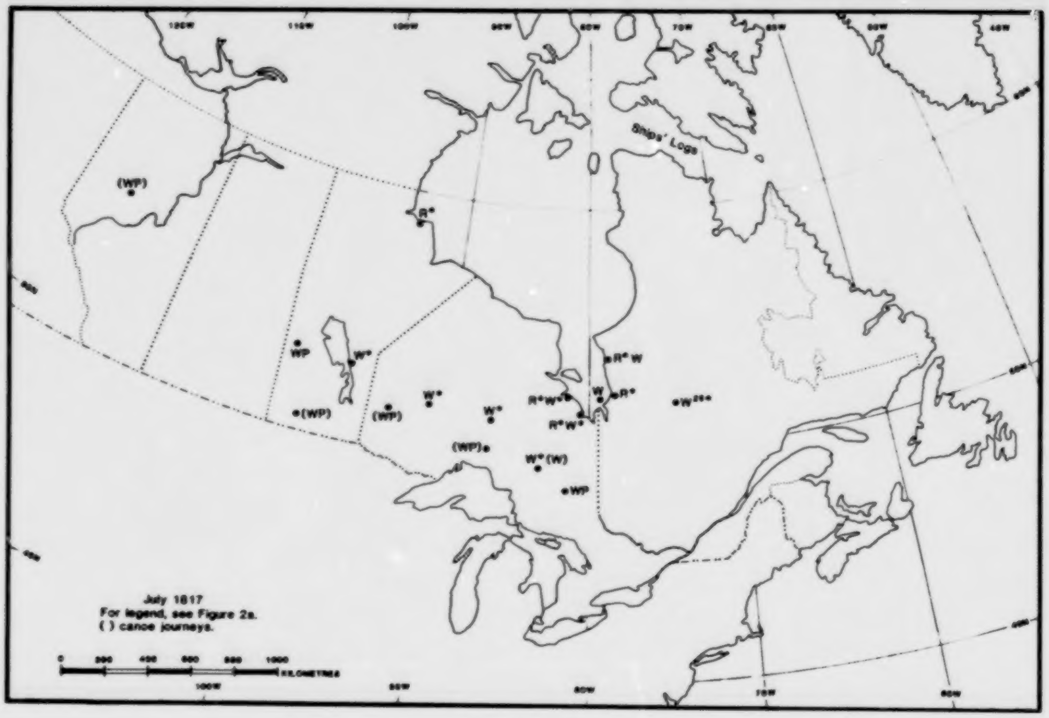

120W
110W
100W
90W
80W
70W
60W
50W
40W
Ships' Logs
(WP)
R*
WP
W*
•(WP)
(WP)
W*
W*
R*W*
R*W*
W
•R*W
•R*
•W²⁺*
(WP)•
W*(W)
•WP
July 1817
For legend, see Figure 2a.
( ) canoe journeys.
0
200
400
600
800
1000
KILOMETRES
100W
90W
80W
70W
60W

FIGURE 2C: *Hudson's Bay Company records, weather data coverage for July 1817.*

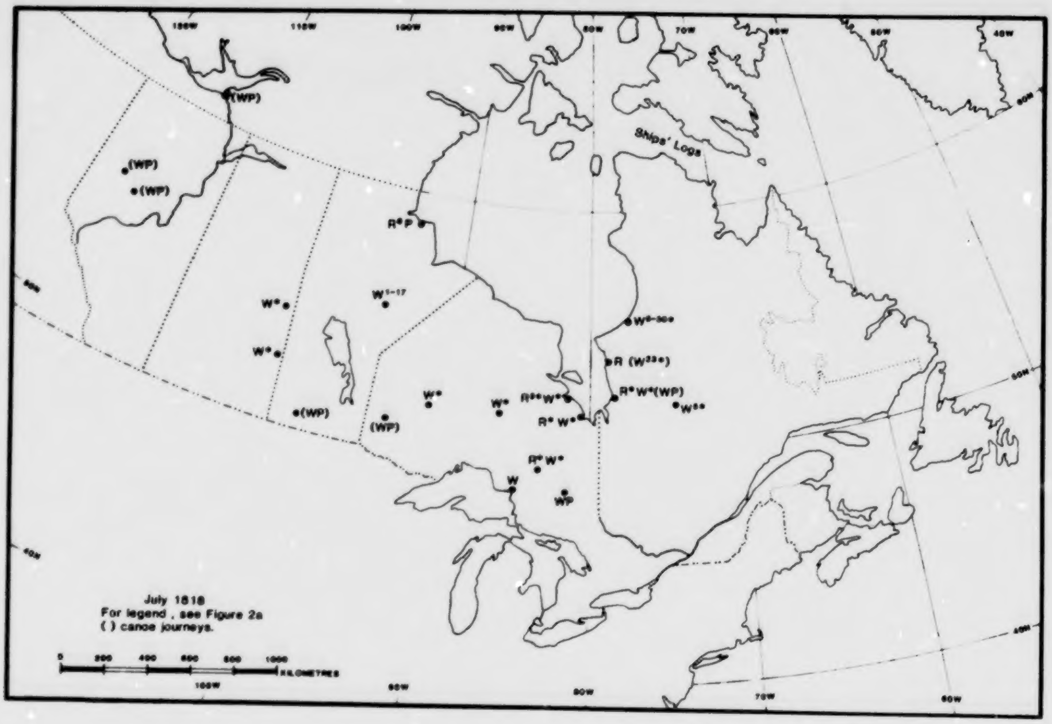

120W
110W
100W
90W
80W
70W
60W
50W
40W
(WP)
(WP)
(WP)
Ships' Logs
R* P
W$^{1-17}$
W*
W*
W^{8-30*}
R (W^{23*})
R* W*(WP)
W^{3*}
(WP)
(WP)
W*
W*
R^{2*} W*
R* W*
R* W*
W
WP
60N
50N
40N
July 1818
For legend , see Figure 2a
( ) canoe journeys.
0
200
400
600
800
1000
KILOMETRES

FIGURE 2D: *Hudson's Bay Company records, weather data coverage for July 1818.*

activity in 1818. In the east, data were logged on journeys exploring the hinterland, from Whale River in 1816, Richmond Gulf in 1818, and from Neoskweskau in 1816 and 1817. There is also the valuable record for each season in the Logs of the ship's inbound from England, from the southwest of Greenland in late June/early July through Hudson Strait and across the Bay to York or Moose Factories. In 1816, ice held the ships in the Strait until early September; in 1817 and 1818, ships crossed the Bay in August.

OTHER DATA SOURCES

At this point, initial steps were taken to check other possible sources of weather information for this period in Canada, and in neighbouring regions of the United States and Greenland, to augment the coverage.

Daily weather information has been obtained for Québec City and Montréal from diaries kept by Dr. Sparks and by John McCord (McGill, 1491-319, 320; cf. Hillaire-Marcel *et al.* 1981), for Halifax from Lady Sherbrooke's diaries, and for Godthaab in western Greenland. These data have been incorporated into the study. A number of Weather Journals are available for the eastern United States at this period, and several have already formed the basis of studies for New England (for example, Milham 1925; Henry 1927; Hoyt 1958; Ludlam 1966; Stommel and Stommel 1979; and Baron and Gordon, this volume). From contacts presently made with archivists in Canada and the United States, it seems that, east of Saskatchewan, sources of weather information also exist (or may well be found) for other key regions. The search continues.

SAMPLE MAP ANALYSIS

Owing to the limited time presently available for trial map analysis, it seemed wiser to choose an extended sequence of days for one season, rather than selecting occasional examples from all three years. In practical terms, this retains the overwhelming advantage of historical continuity, once an initial situation has been analyzed and a foothold established. In real terms, from a climatological viewpoint, a sequence offers a better opportunity to assess the usefulness of such a study. Three criteria were invoked in the choice of period: (1) that there should already be some knowledge of the general situation--i.e., it is useful to proceed from the known to the unknown in setting up and testing a methodology; (2) for similar reasons, that it be a period with relatively good data coverage; and (3) that the sequence should have climatic interest.

A decision was first made to restrict the choice to the 1816 season. For the 123 days from May 1 to August 31, 1816, all the direct and proxy weather information was plotted on daily base maps. These day-by-day summaries provide the basic working sheets and were studied carefully. A comparison of the charts for May and August 1816 indicates the upper and lower limits of the Hudson's Bay Company database for the three seasons (see for example, Figures 3A, B). The run of days from June 1 to June 17, 1816 was finally chosen for analysis.

Method of Analysis

Using transparent overlays, the series of data for morning and evening hours were transposed from the base map onto separate charts. Each chart was then analyzed over the base map, which provided the necessary background information as well as intermediate history, and in conjunction with previous maps. Areas of cloud and precipitation were shaded in, the temperature and wind fields studied, and zones of maximum gradient and wind shear noted. Pressure tendencies were available for Québec City, Godthaab and for New Haven (reproduced for June 1-13 in Stommel and Stommel 1979). Frontal zones were tentatively indicated, and an attempt was then made to sketch in the pressure pattern, bearing in mind surface wind speed, the nature of the surface and the most likely direction and speed of movement of the frontal systems.

The latter is critical, and was given careful consideration by Loader in the study for the Armada Period, July to October 1588 by Douglas *et al.* (1978, pp. 6,9). Guidelines were finally obtained from Palmén's (1928) investigation. In brief: *for open frontal waves*, the speed of displacement over the land is generally greater than the surface wind measured in the warm air, often by 50% to 80%. In contrast, over the open sea, it may be only 75% to 90% of the observed surface wind speed in the warm air; *with young (warm-sector) cyclones up to the early stages of occlusion*, the speed of movement is usually less than that of an open wave. While warm fronts move at approximately the same rate as that for the low centres, cold fronts tend to move more rapidly (Petterssen 1956): *in the case of old occluded cyclones*, the speed of displacement decreases rapidly as the system ages, unless it is regenerated by a fresh inflow of warm or cold air.

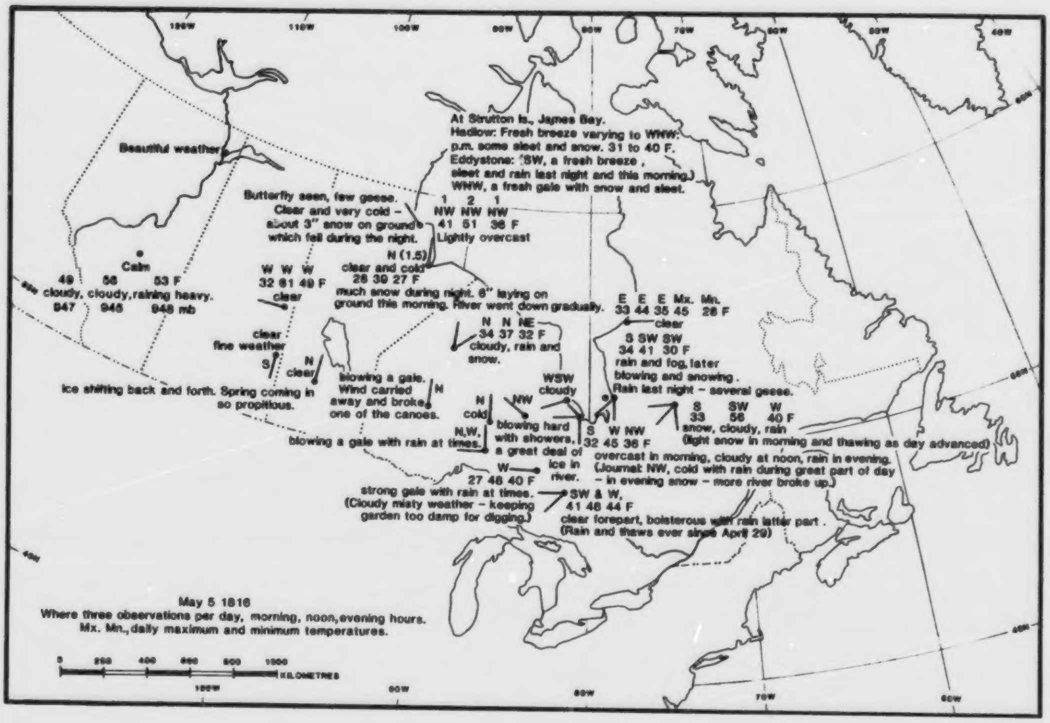

120W
110W
100W
90W
80W
70W
60W
50W
40W
60N
50N
40N
At Strutton Is., James Bay.
Hadlow: Fresh breeze varying to WNW:
p.m. some sleet and snow. 31 to 40 F.
Eddystone: (SW, a fresh breeze ,
sleet and rain last night and this morning.)
WNW, a fresh gale with snow and sleet.
Beautiful weather
Butterfly seen, few geese.
Clear and very cold -
about 3" snow on ground
which fell during the night.
1 2 1
NW NW NW
41 51 36 F
Lightly overcast
N (1.5)
clear and cold
28 30 27 F
much snow during night. 6" laying on
ground this morning. River went down gradually.
Calm
49 56 53 F
cloudy, cloudy, raining heavy.
947 946 948 mb
W W W
32 61 49 F
clear
clear
fine weather
S
N
clear
ice shifting back and forth. Spring coming in
so propitious.
blowing a gale.
Wind carried
away and broke
one of the canoes.
N
N N NE
34 37 32 F
cloudy, rain and
snow.
E E E Mx. Mn.
33 44 35 45 28 F
clear
S SW SW
34 41 30 F
rain and fog, later
blowing and snowing .
Rain last night - several geese.
WSW
cloudy
N
cold
NW
N.W.
blowing a gale with rain at times.
blowing hard
with showers,
a great deal of
ice in
river.
S W NW
32 45 36 F
overcast in morning, cloudy at noon, rain in evening.
(Journal: NW, cold with rain during great part of day
- in evening snow - more river broke up.)
S SW W
33 56 40 F
snow, cloudy, rain
(light snow in morning and thawing as day advanced)
W
27 48 40 F
strong gale with rain at times.
(Cloudy misty weather - keeping
garden too damp for digging.)
SW & W,
41 48 44 F
clear forepart, boisterous with rain latter part .
(Rain and thaws ever since April 29)
May 5 1816
Where three observations per day, morning, noon, evening hours.
Mx. Mn., daily maximum and minimum temperatures.
0 200 400 600 800 1000 KILOMETRES

FIGURE 3A: *Daily weather summaries, based on the Hudson's Bay Company records for May 5, 1816 (sample showing maximum coverage).*

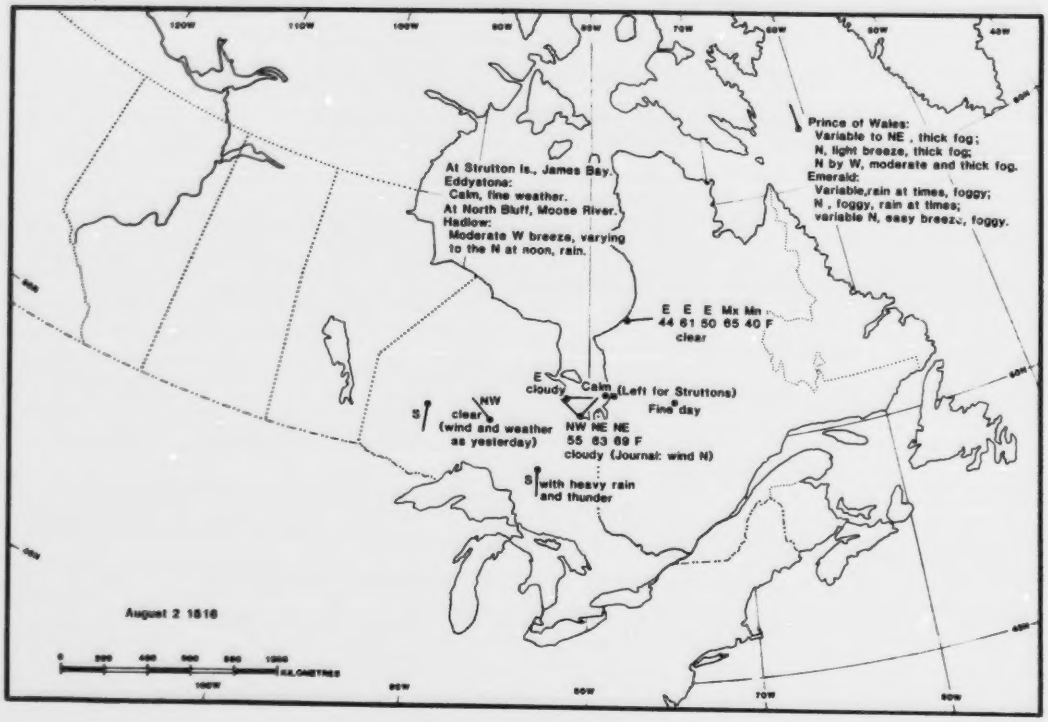

120W
110W
100W
90W
80W
70W
60W
50W
40W
60N
50N
40N
At Strutton Is., James Bay.
Eddystone:
Calm, fine weather.
At North Bluff, Moose River.
Hadlow:
Moderate W breeze, varying
to the N at noon, rain.
Prince of Wales:
Variable to NE , thick fog;
N, light breeze, thick fog;
N by W, moderate and thick fog.
Emerald:
Variable,rain at times, foggy;
N , foggy, rain at times;
variable N, easy breeze, foggy.
E E E Mx Mn
44 61 50 65 40 F
clear
E
cloudy
Calm
(Left for Struttons)
Fine day
S
NW
clear
(wind and weather
as yesterday)
NW NE NE
55 63 69 F
cloudy (Journal: wind N)
S
with heavy rain
and thunder
August 2 1816
0 200 400 600 800 1000
KILOMETRES

FIGURE 3B: *Daily weather summaries, based on the Hudson's Bay Company records for August 2, 1816 (sample showing minimum coverage).*

As an aid to analysis, wherever Registers were available, temperature curves for morning and evening observations were drawn up by month with winds, cloud and precipitation added, to give a visual display of the sequence of weather (Figure 4). Pressure traces were also obtained for Québec City, Godthaab and New Haven. In order to look more closely at the relationship between the rise and fall of temperature and pressure, these two sets of data were plotted for Albany for the summer of 1817, and for Eastmain in 1818.

Degree of Resolution

Just by studying the sequence of daily summary charts and the station time series (Figures 3, 4), insights can immediately be gained as to the general nature of the flow patterns and their changes. When the flow situations are relatively simple in this part of Canada, as on June 6, 1816, a generalized pressure distribution can readily be sketched in. To go further, to locate systems and plot their daily movement with even a limited degree of accuracy is not always easy using the Hudson's Bay Company records alone. Too many options may be left open. But it was interesting to see how dramatically the addition of data for Montréal, Québec City, Godthaab, New Haven and Waltham, Massachusssets (reproduced for June 1816 in Ludlam 1966) reduced these options and increased the degree of confidence. Although a synoptic map series for Canada at this early period cannot approach the high degree of confidence of that of Kington's (1975,1980), owing to the rarity of pressure readings and paucity of wind speeds, the series can be increasingly refined by successive approximation as further data are acquired. In addition, as Douglas and Lamb (1979) have shown, such new information serves as a useful check on the confindence limits of the analysis to date. Several regional climatic studies of New England for this period (including that of Ludlam 1966) serve a similar dual purpose. In the light of this discussion, the maps presented below offer an early approximation.

SYNOPTIC SITUATION, JUNE 1-17, 1816

Weather maps for June 4,5,9,10,11 and 16 (Figure 5A-F) illustrate some of the significant changes in the low-level circulation patterns during this 17-day period. Figure 6 summarizes the movement of the surface systems.

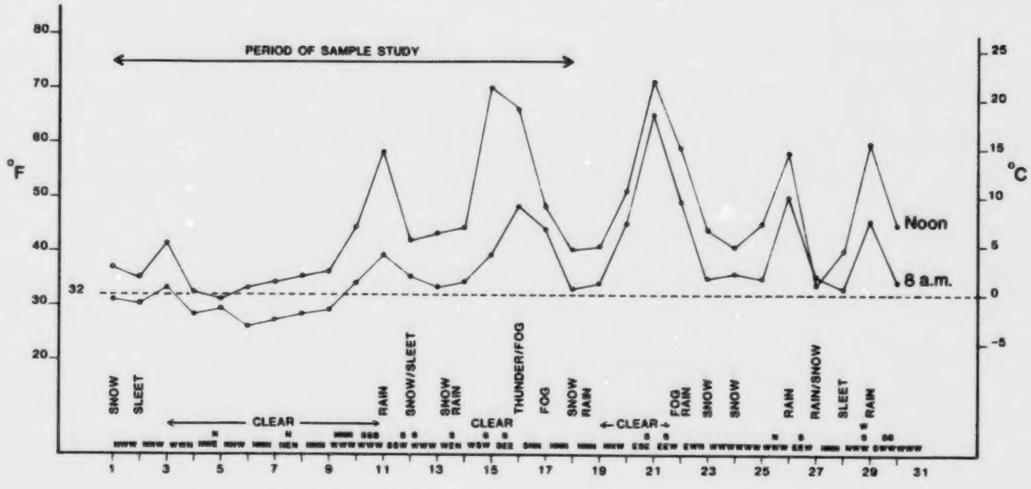


FIGURE 4: *Whale River, June 1816: sequence of daily temperature (8 am, noon), winds and weather.*

A tentative description of the synoptic events over central and eastern Canada from June 4 to 11 was put forward in a recent study (Wilson 1983B), based on weather reports for sites on Hudson/James Bay and Ludlam's reconstruction for New England. It was suggested that the sequence represented a classic example of a springtime Hudson Bay high, as described by Johnson (1948). Typically, cold surface anticyclones entering the Hudson Bay region intensify under the influence of a warm upper-air ridge over east and central North America and eventually move out to the southeast. Such a circulation pattern may persist for five days or more. It is usually associated with a below-normal zonal index, and a shorter wave-length than normal, with troughs over the western United States and off the east coast.

From June 2, a depression moved from the southwest, to the north of the Great Lakes and into New Québec/Labrador. In its wake, cold air flowed south from northern Canada. By the evening of June 5, the northeastward movement of the low had slackened, and from June 6 to 9, the depression stalled (or drifted only very slowly) over New Québec/Labrador. The system, now occluding, trailed a deep trough down over eastern Canada and the United States and out into the Atlantic. During this time, a high pressure cell was building over Hudson Bay, where the ice cover was unusually complete and compacted. On June 9, this anticyclone began to move out rapidly to the southeast and was centred off the coast of New England by the evening of June 11.

As the high shifted to the southeast, southerly flow around its western flank brought warmer weather to parts of eastern Canada and the Bay region. Another depression approaching from the west on June 9 dipped southward across the Bay to stall and fill over northern Québec during June 13-14. This in turn was followed by a low on a rather more northerly track across the Bay from June 13 to 17. At this time, high pressure built up over the eastern half of Canada, bringing several days of very warm, fine, summer weather. On the evening of June 16, a trailing cold front from the low now over the east coast of the Bay triggered a series of heavy thunderstorms in the warm unstable airmass over northern Ontario and the coast of James Bay.

The synoptic analysis for June 1-17, 1816 indicates a period marked by sharp contrasts in temperature and weather, such as accompany strong meridional circulation patterns. It also appears to confirm the prolonged intensification, then rapid southeastward shift of a Hudson Bay anticyclone from June 4 to 11. Johnson (1948) found that such situations in Canada were preceded by blocking over the North Atlantic and western Europe. This is in keeping with the results of the recent climatological study of blocking by Treidl *et al.*

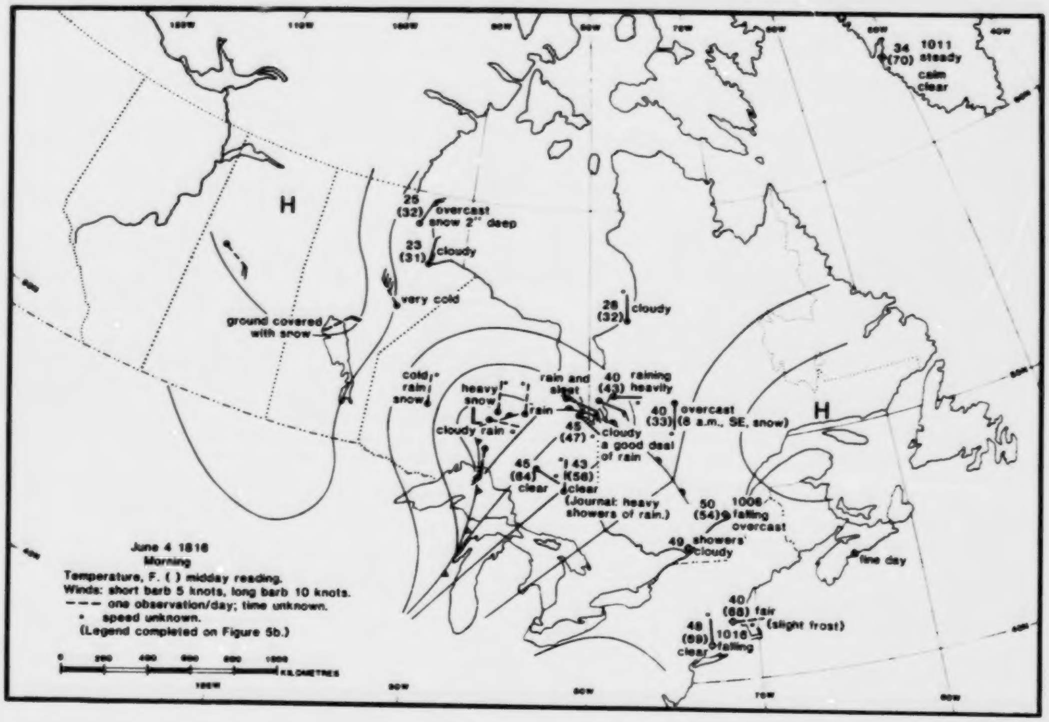

120W
110W
100W
90W
80W
70W
60W
50W
40W
H
25
(32) overcast
snow 2'' deep
23
(31) cloudy
very cold
ground covered
with snow
28
(32) cloudy
34 1011
(70) steady
calm
clear
cold
rain
snow
heavy
snow
L
cloudy rain
rain
rain and
sleet
40 raining
(43) heavily
45
(47)
cloudy
a good deal
of rain
40 overcast
(33) (8 a.m., SE, snow)
H
45
(64)
clear
43
(56)
clear
(Journal: heavy
showers of rain.)
50 1006
(54) falling
overcast
49 showers'
cloudy
fine day
40
(68) fair
(slight frost)
48
(69) 1016
clear falling
June 4 1816
Morning
Temperature, F. ( ) midday reading.
Winds: short barb 5 knots, long barb 10 knots.
— — — one observation/day; time unknown.
• speed unknown.
(Legend completed on Figure 5b.)
0
200
400
600
800
1000
KILOMETRES
100W
90W
80W
70W
60W

FIGURE 5A: *Surface weather map: morning, June 4, 1816.*

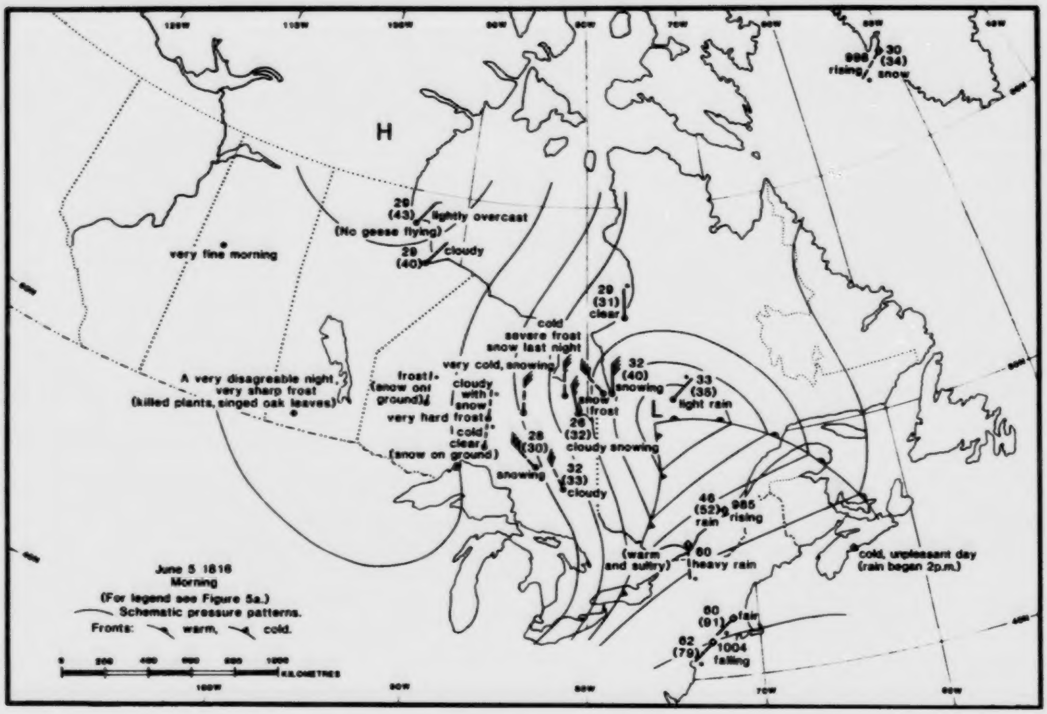

120W
110W
100W
90W
80W
70W
60W
50W
40W
H
very fine morning
29
(43)
lightly overcast
(No geese flying)
29
(40)
cloudy
29
(31)
clear
cold
severe frost
snow last night
very cold, snowing
A very disagreable night,
very sharp frost
(killed plants, singed oak leaves)
frost
(snow on ground)
cloudy
with
snow
very hard frost
cold
clear
(snow on ground)
28
(30)
snowing
32
(33)
cloudy
snow
frost
26
(32)
cloudy snowing
32
(40)
snowing
33
(35)
light rain
L
996
rising
30
(34)
snow
46
(52)
rain
985
rising
(warm
and sultry)
60
heavy rain
cold, unpleasant day
(rain began 2p.m.)
60
(91)
fair
62
(79)
1004
falling
June 5 1816
Morning
(For legend see Figure 5a.)
Schematic pressure patterns.
Fronts: warm, cold.
0
200
400
600
800
1000
KILOMETRES
100W
90W
80W
70W
60W

FIGURE 5B: *Surface weather map: morning, June 5, 1816.*

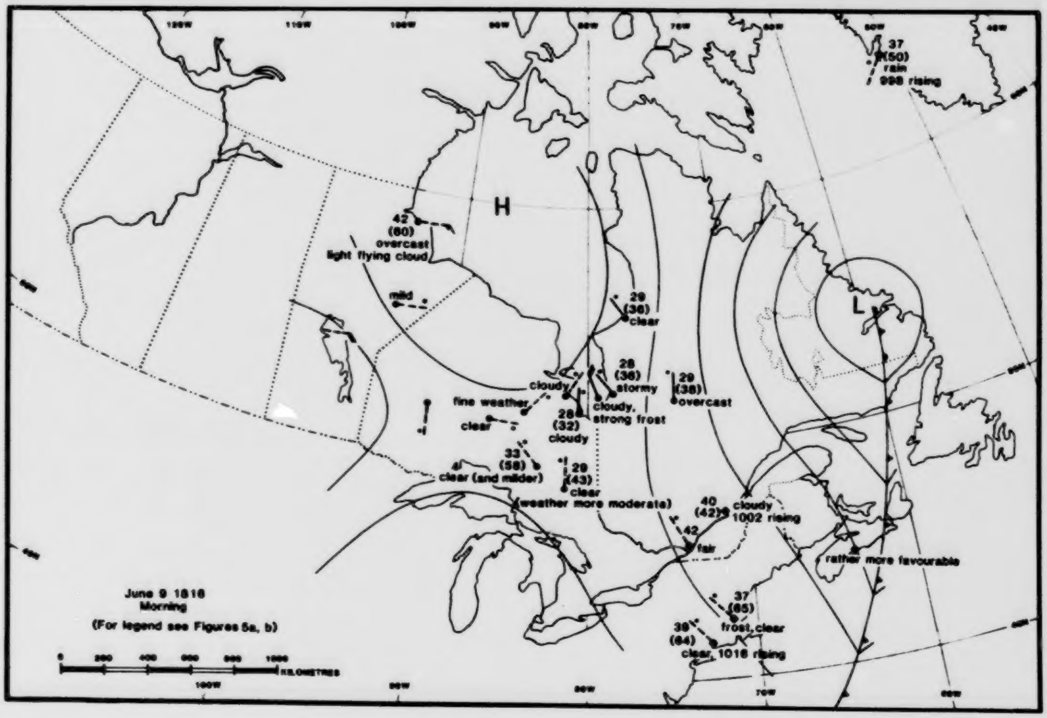

120W
110W
100W
90W
80W
70W
60W
50W
40W
37
(50)
rain
998 rising
H
42
(60)
overcast
light flying cloud
mild
29
(36)
clear
L
28
(36)
stormy
29
(38)
overcast
cloudy
fine weather
clear
28
(32)
cloudy
cloudy,
strong frost
33
(58)
clear (and milder)
29
(43)
clear
(weather more moderate)
40
(42)
cloudy
1002 rising
42
fair
rather more favourable
37
(65)
frost, clear
39
(64)
clear, 1016 rising
June 9 1816
Morning
(For legend see Figures 5a, b)
0
200
400
600
800
1000
KILOMETRES
100W
90W
80W
70W
60W

FIGURE 5C: *Surface weather map: morning, June 9, 1816.*

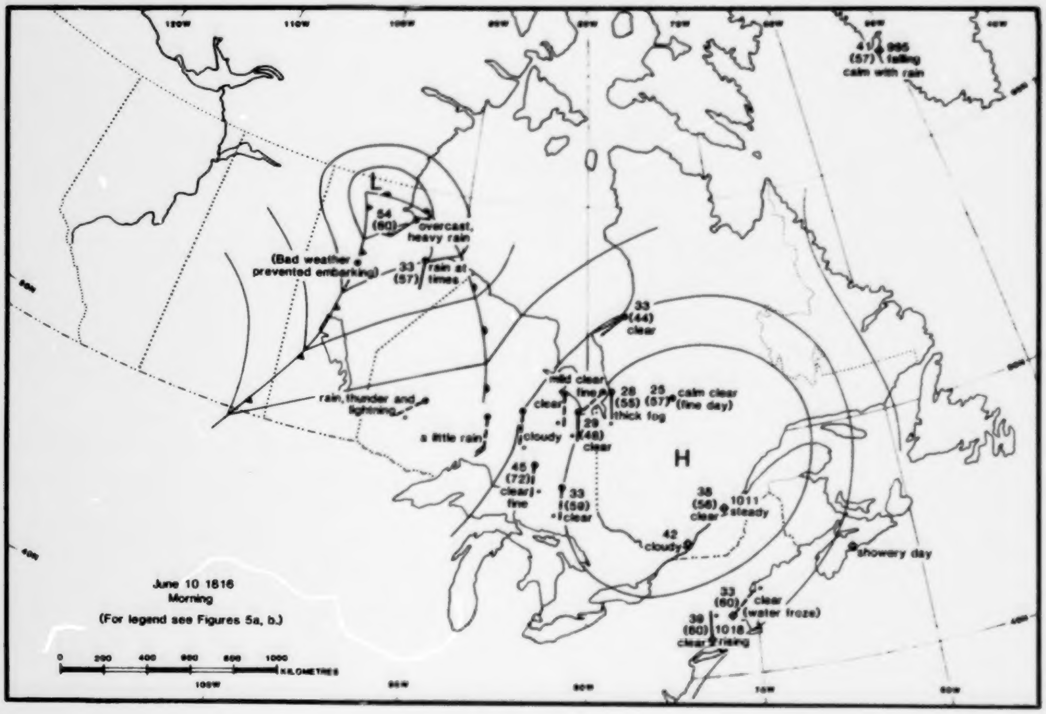

120W
110W
100W
90W
80W
70W
60W
50W
40W
41
(57)
985
falling
calm with rain
L
54
(60)
overcast,
heavy rain
(Bad weather
prevented embarking)
33
(57)
rain at
times
33
(44)
clear
rain,thunder and
lightning
a little rain
mild clear
fine
clear
cloudy
28
(55)
thick fog
25
(57)
calm clear
(fine day)
29
(48)
clear
H
45
(72)
clear
fine
33
(59)
clear
38
(56)
clear
1011
steady
42
cloudy
showery day
33
(60)
clear
(water froze)
39
(60)
clear
1018
rising
June 10 1816
Morning
(For legend see Figures 5a, b.)
0
200
400
600
800
1000
KILOMETRES
100W
90W
80W
70W
60W
40N
50N

FIGURE 5D: *Surface weather map: morning, June 10, 1816.*

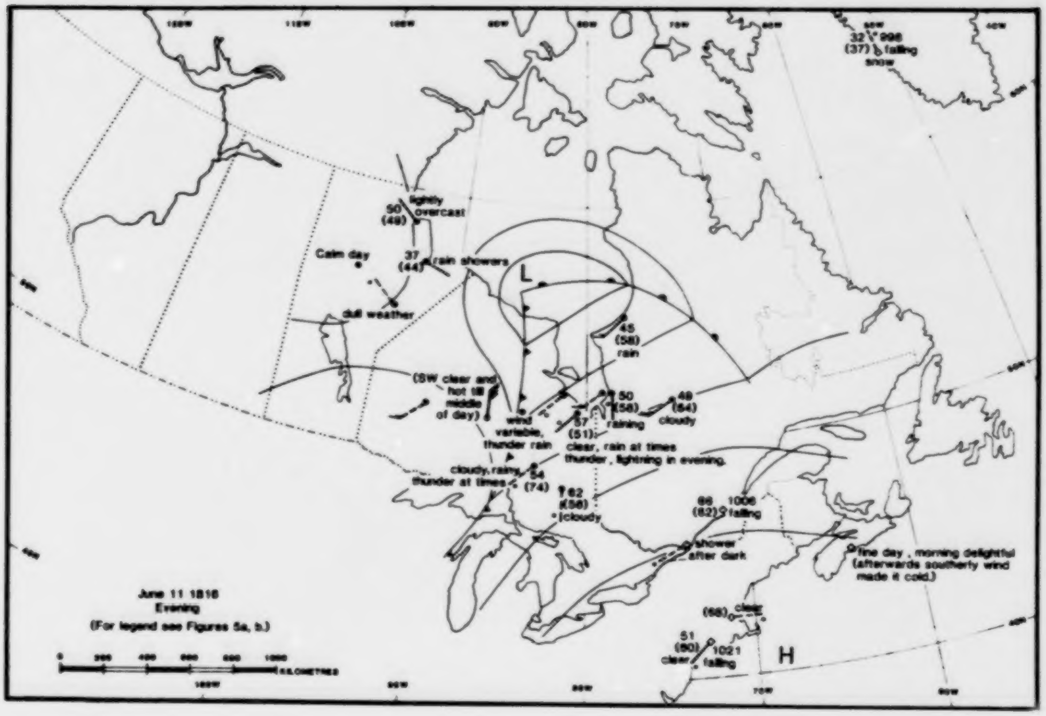

120W
110W
100W
90W
80W
70W
60W
50W
40W
32
998
(37)
falling
snow
50
(49)
lightly
overcast
Calm day
37
(44)
rain showers
dull weather
L
45
(58)
rain
(SW clear and
hot till
middle
of day)
wind
variable,
thunder rain
50
(58)
raining
49
(64)
cloudy
57
(51)
clear, rain at times
thunder, lightning in evening.
cloudy, rainy,
thunder at times
54
(74)
62
(58)
cloudy
66
(62)
1006
falling
shower
after dark
fine day, morning delightful
(afterwards southerly wind
made it cold.)
(66)
clear
51
(60)
clear
1021
falling
H
June 11 1816
Evening
(For legend see Figures 5a, b.)
0
200
400
600
800
1000
KILOMETRES
100W
90W
80W
70W
60W

FIGURE 5E: *Surface weather map: evening, June 11, 1816.*

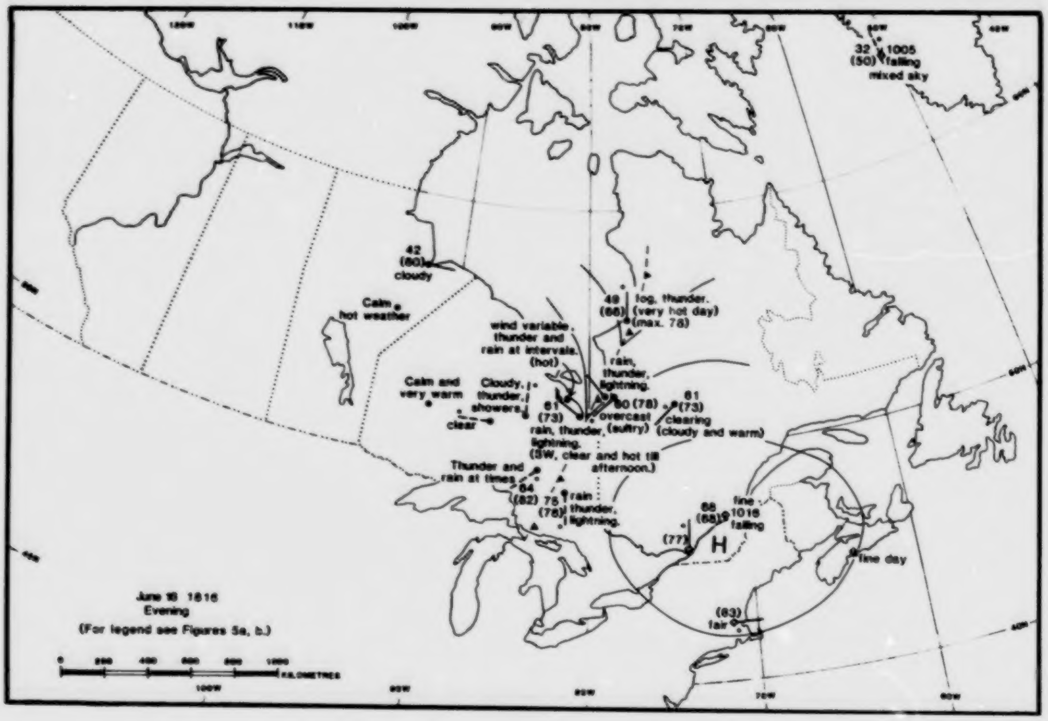

120W
110W
100W
90W
80W
70W
60W
50W
40W
32
(50)
1005
falling
mixed sky
42
(60)
cloudy
Calm
hot weather
49
(66)
fog, thunder.
(very hot day)
(max. 78)
wind variable,
thunder and
rain at intervals.
(hot)
rain,
thunder,
lightning.
Calm and
very warm
Cloudy,
thunder,
showers
clear
61
(73)
rain, thunder,
lightning.
(SW, clear and hot till
afternoon.)
60 (78)
overcast
(sultry)
61
(73)
clearing
(cloudy and warm)
Thunder and
rain at times
64
(82)
75
(78)
rain
thunder,
lightning.
66
(68)
fine
1016
falling
(77)
H
fine day
(63)
fair
June 16 1816
Evening
(For legend see Figures 5a, b.)
0
200
400
600
800
1000
KILOMETRES
100W
90W
80W
70W
60W

FIGURE 5F: *Surface weather map: evening, June 16, 1816.*

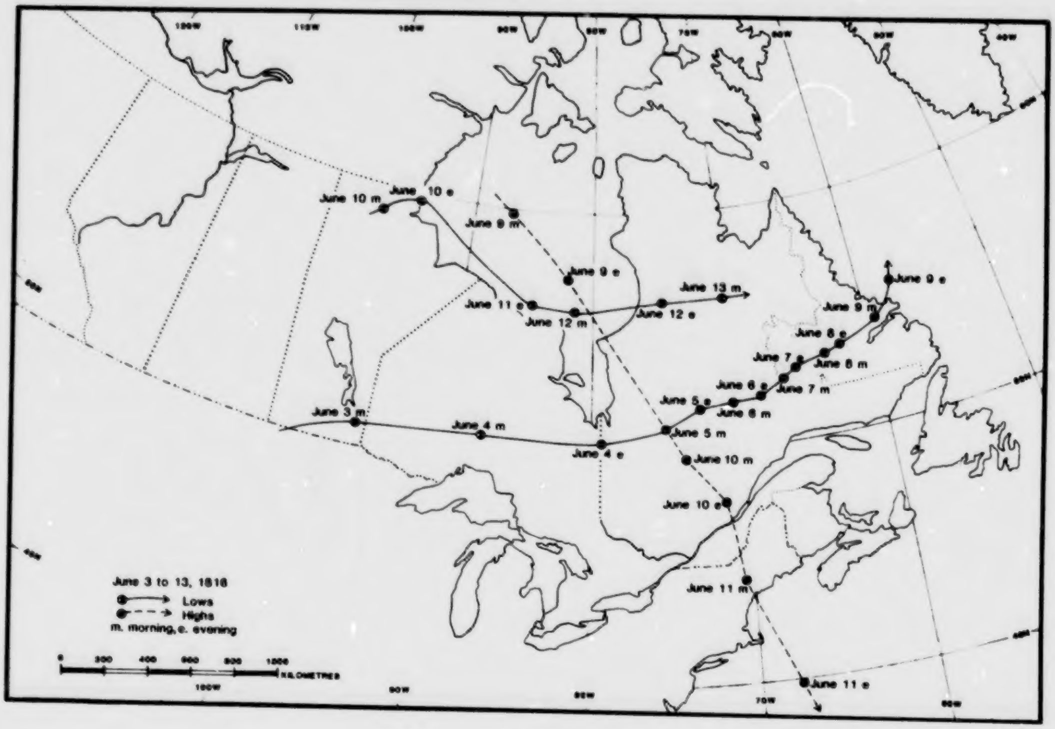

120W
110W
100W
90W
80W
70W
60W
50W
40W
June 10 m
June 10 e
June 9 m
June 9 e
June 11 e
June 12 m
June 12 e
June 13 m
June 9 e
June 9 m
June 8 e
June 7 e
June 8 m
June 6 e
June 7 m
June 5 e
June 6 m
June 3 m
June 4 m
June 4 e
June 5 m
June 10 m
June 10 e
June 11 m
June 11 e
June 3 to 13, 1816
Lows
Highs
m. morning, e. evening
0
200
400
600
800
1000
KILOMETRES
100W
90W
80W
70W
60W

FIGURE 6: *Trajectories of the surface systems (June 3-13, 1816).*

(1981). Possibly these Hudson Bay highs may have occurred with unusual frequency and persistence in the spring and summer of 1816. This view is reflected in the circulation patterns for Europe and the North Atlantic reconstructed for July 1816 by Lamb and Johnson (1966), and in the dry growing season experienced in southern Manitoba at this time (Allsopp 1977).

CONCLUSION

The findings of this study, suggest that a useful series of daily weather maps could be obtained for this early period, although for best results the Hudson's Bay Company weather database needs to be supplemented by similar information for key areas from other sources. The ultimate value of such a series is primarily as an indicator of the general circulation of the atmosphere, based on the known relationships between the surface pressure systems and the long-wave patterns of the upper troposphere--between the location of the surface highs and lows and the corresponding ridges, troughs and closed centres aloft, the trajectories of surface lows and the alignment and speed of the jetstreams. A map study of this kind should provoke some answers to the fascinating questions concerning the differences in the nature and frequency of the general circulation patterns between the summers of 1816, 1817, 1818, and permit comparison with those of recent seasons.

ACKNOWLEDGEMENTS

I am grateful to the National Museum of Natural Sciences Climatic Change Project (Coordinator, Dr. C.R. Harington) for financial support for this study, and for encouragement. I thank also: Mrs. Shirlee Anne Smith and her staff at the Hudson's Bay Company Archives for help in locating some of the Posts; Ms. Sandra M. Haycock, Public Records Archivist, Public Archives of Nova Scotia, for providing material from Lady Sherbrooke's diaries; Dr. Karl Andersen and his staff at the Meteorologisk Institut in Copenhagen for making available copies of the data for Godthaab; Mr. Mike Chenowith of Fulton, Missouri, for letting me know of the Godthaab source and of some other possible data sources in the United States; and Mrs. Valerie Moore, who kindly typed the manuscript. Edward Hearn, University of Ottawa, drafted the figures under contract to the National Museum of Natural Sciences.

REFERENCES

Allsopp, T.R. 1977. Agricultural weather in the Red River basin of southern Manitoba over the period 1800 to 1975. Atmospheric Environment Service (Downsview) CLI-3-77:1-28.

Atmospheric Environment Service. 1970. Beaufort wind scale. DS #5-70:1-4.

Baron, W.R., and G.A. Gordon. (This volume). A reconstruction of New England climate using historical materials, 1620-1980.

Douglas, K.S., H.H. Lamb, and C. Loader. 1978. A meteorological study of July to October 1588: the Spanish Armada storms. University of East Anglia, Norwich, Climatic Research Unit Research Publication No. 6:1-16.

Douglas, K.S., and H.H. Lamb. 1979. Weather observations and a tentative meteorological analysis of the period May to July 1588. University of East Anglia, Norwich, Supplement to Climatic Research Unit Research Publication No. 6:1-4.

Henry, A.J. 1927. Abnormal summers in the United States. Monthly Weather Review 55: 349-353.

Hillaire-Marcel, C., S. Occhietti, L. Marchand, and R. Rajewicz. 1981. Analysis of recent climatic changes in Québec: some preliminary data. In: Climatic Change in Canada 2, Edited by: C.R. Harington. Syllogeus No. 33:28-47.

Hoyt, J.B. 1958. The cold summer of 1816. Annals of the Association of American Geographers 48:118-131.

Johnson, C.B. 1948. Anticyclogenesis in eastern Canada during spring. Bulletin of the American Meteorological Society 29:45-55.

Kington, J.A. 1975. Daily synoptic weather maps from the 1780s: a research project of synoptic climatology. Meteorological Magazine 104:33-52.

_______. 1980. Daily weather mapping from 1781: a detailed synoptic examination of weather and climate during the decade leading up to the French Revolution. Climatic Change 3:7-36.

Lamb, H.H., and A.I. Johnson. 1966. Secular variations of the atmospheric circulation since 1750. Meteorological Office (London), Geophysical Memoirs No. 110:1-125.

Ludlam, D. 1966. Early American winters 1604-1820. American Meteorological Society. pp. 190-194.

Milham, W.I. 1925. The year 1816 - the causes of abnormalities. Monthly Weather Review 52: 563-570.

Oliver, J., and J.A. Kington. 1970. The usefulness of Ships' Log-books in the synoptic analysis of past climates. Weather 25:520-528.

Palmén, E. 1928. Zur Frage der Fortpflanzungsgeschwindigkeit der Zyklonen. Meteorologische Zeitschrift 45:96-99.

Petterssen, S. 1941. Introduction to meteorology. McGraw-Hill, New York. 236 pp.

_______. 1956. Weather analysis and forecasting. McGraw-Hill, New York, Volume I, 428 pp. Volume II, 266 pp.

_______. 1958. Introduction to meteorology. Second edition. McGraw-Hill, New York. 327 pp.

Saucier, W.J. 1955. Principles of meteorological analysis. University of Chicago Press Chicago. 438 pp.

Stommel, H., and E. Stommel. 1979. The year without a summer. Scientific American 240: 176-186.

Treidl, R.A., E.C. Birch, and P. Sajecki. 1981. Blocking action in the northern hemisphere: a climatological study. Atmosphere-Ocean 19:1-23.

Wilson, C. 1982. The summer season along the east coast of Hudson Bay during the nineteenth century. Part I. General introduction; climatic controls: calibration of the instrumental temperature data, 1814 to 1821. Canadian Climate Centre Report No. 82-4: 1-223.

______. 1983A. Some aspects of the calibration of early Canadian temperature records in the Hudson's Bay Company Archives: a case study for the summer season, eastern Hudson/James Bay, 1814 to 1821. In: Climatic Change in Canada 3. Edited by: C.R. Harington. Syllogeus No. 49:144-202.

______. 1983B. The Little Ice Age on eastern Hudson Bay: summers at Great Whale, Fort George, Eastmain, 1814-1821. Canadian Climate Centre Report No. 83-9:1-145.

DATA SOURCES

Hudson's Bay Company Archives (HBC), Winnipeg, Manitoba.

McGill University Archives, Montrêal, P.Q.

Meteorologisk Institut, København, Denmark (Data for Godthaab, Greenland.)

Ludlam, D. 1966. p. 193. Reproduction of a letter to the Sentinel, July 1, 1816, containing a weather diary for Waltham, Massachussetts, June 1816.

Stommel, H., and E. Stommel. 1979. p. 178. Reproduction of the Meteorological Journal (June 1-13, 1816) for New Haven. (Original manuscript, Yale University.)

A DRAMATIC CHANGE IN THE GENERAL CIRCULATION ON THE WEST COAST OF HUDSON BAY IN 1760 A.D.: SYNOPTIC EVIDENCE BASED ON HISTORIC RECORDS

Timothy Ball[1]

INTRODUCTION

The Daily Journals maintained by employees of the Hudson's Bay Company Posts contain daily references to the weather. These references were transferred by means of a coding system in order to produce long-term sequences of weather events at York Factory and Churchill Factory. An overview and preliminary analysis of this work was published previously (Catchpole and Ball 1981). More detailed analysis is presently in progress - particularly of the instrumental records commencing in 1768 at Churchill.

A year-by-year synopsis of the climatic variables appears to indicate that there was a significant shift in the general circulation, probably as a result of a change in the mean summer position of the Arctic Front. This shift is particularly apparent in a change in the frequency of precipitation events at York Factory, but it is also evident in other climatic variables.

Lamb (1977, p. 463) points out that there was an advance of European Alpine glaciers during the period 1780-1850. He suggests that this was due to an increase in precipitation rather than a decrease in tempertures. An expansion of the circumpolar vortex resulted in:

> "... displacements of the most frequented depression tracks to somewhat lower latitudes than in other epochs, frequent wetness of the summers in much of Europe." (Lamb 1977, p.466).

A study of the pressure patterns in the latitudes of the westerlies indicated that,

> "The southern position of the low pressure trough right across the eastern Atlantic and European longitudes on the July map for the 1690s was more or less matched in the 1760s and 1810-19." (Lamb, 1977, p.490).

Unfortunately there are few synoptic studies of the southern shores of Hudson Bay, but there are several peripheral studies. There have been quite extensive climatological studies of the Canadian Arctic Islands - particularly Baffin Island, for example the works of Brinkmann and Barry (1972); Barry *et al.* (1975); Locke and Locke (1977); and Alt (1978). These have evolved because of the availability of brief but detailed modern records.

---

[1] Department of Geography, University of Winnipeg, Winnipeg, Manitoba R3B 2E9

A study by Minns (1970) was based on the air mass and frontal climatology of Bryson. Minns hypothesized that it is possible to estimate the presence of an air-mass type at a certain location given a particular temperature. This temperature is seen as being conditionally dependent upon wind, precipitation, and cloud conditions, and therefore itself can be estimated by knowledge of these meteorological elements. Using these probabilities, a stochastic model was created reaching the conclusion that a weakened westerly vortex led to an increased intrusion of Arctic air during the nineteenth century.

Certainly north winds experienced along the southern shores of Hudson Bay would be bringing air from Arctic latitudes, and this would probably result in cooler conditions than exist with winds from other directions. The difficulty is that north winds can be the result of variety of pressure patterns, each of which can result from varying atmospheric situations. Barry *et al.* (1975) identified 12 synoptic conditions affecting different regions of central Canada. Their diagrams clearly show that northerly-component winds occur under several of these situations in the southern Hudson Bay region.

Barry *et al.* (1975) attribute cooler summers and increased passage of cyclonic systems with "...a westward displacement of the mean 700mb trough over eastern North America,...". These results are interesting when linked with the results of Locke and Locke, who note that,

> "The depression of glaciation thresholds and equilibrium-line altitudes of about 400m along the coast of Baffin Bay may indicate increased precipation during the Little Ice Age due to an increase in frequency of low pressure centered over Davis Strait and Baffin Bay." (Locke and Locke 1977, p.299)

Combining these ideas, it would appear that a westward shift of the 700mb trough is associated with a shorter wavelength of the stationary long-wave pattern that may result in a low zonal index and associated meridional conditions. Conversely, an eastward shift may result in a longer wavelength, high zonal index, and more latitudinal flow.

Dey (1973) examined the synoptic conditions that occur during summer dry spells in the Canadian Prairies, concluding:

> "...there was a general eastward migration of all pressure systems and that the north Pacific high tended to extend north eastward from their normal positions, whereas the Hudson Bay low moved eastward during dry spells. (Dey 1973, p.169)

At the 700mb level, he observed an eastward displacement of the high pressure ridge that formed on the lee side of the Rocky Mountains. This would probably result in an eastward displacement of the 700mb trough that, normally, is located in eastern North America.

Putting the findings of Dey (1973) together with those of Barry *et al.* (1975), and Locke and Locke (1977), suggests to me that eastward movement of the north Pacific high and eastward movement of the 700mb trough of eastern North America result in below normal precipitation in the Canadian Prairies and above average precipitation in the southern region of Hudson Bay. The cause of this increase would seem to be the location of the Rossby waves in a low zonal index, associated meridional flow, and an increase in the frequency of cyclonic systems tracking through the western and southern portion of Hudson Bay.

Bryson (1966) and Barry (1967) have both concluded that there is a close correlation between the mean summer position of the Arctic Front and the boreal forest-Tundra boundary. Barry finds this to be particularly true"...in northwest Canada where the forest-tundra boundary is extremely sharp, but suggests that the correlation may not be applicable in Labrador-Ungava." (Barry 1967, p.90). The treeline, which runs in an east-west line in the Churchill region, is quite clearly defined and seems to reflect a relatively clear delineation of the position of the Arctic Front. Although York Factory and Churchill are only two degrees of latitude apart (approximately 195km), the former is well within the forested zone, while Churchill is at the edge. Mitchell (1973) provides an estimated position of the treeline that supports the inclusion of York Factory within the forested zone.

The implication of these facts is that Churchill would be more influenced by Arctic airmasses than York Factory. If the treeline at Churchill is coincident with the mean summer position of the Arctic Front, then it is presumably under tundra conditions at least 50% of the summer. Similarly, York Factory, being south of that line, would experience more southerly flows and boreal forest conditions for most of the summer months. If there were changes in the circumpolar vortex, and therefore in the mean summer position of the Arctic Front, they would be manifested in changes of the climatic variables at stations close to the Front. Data generated from analysis of the Daily Journals maintained by the Hudson's Bay Company seem to indicate a shift in the mean summer position of the Arctic Front about 1760 A.D.

A synoptic study of climatic conditions in the Canadian Arctic (Fletcher and Young 1976, p.43) presents a map showing isolines connecting points of equal rainfall days. One of these lines runs between York Factory and Churchill indicating that Churchill has fewer rainfall days than York Factory in the modern record. This is the situation that would be expected if the mean summer position of the Arctic Front lay between the two situations, thus giving York Factory a greater influx of unstable southerly air in the summer months and therefore more convective rainfall.

Figure 1 shows the total number of days on which some rainfall event occurred for each year from 1715 to 1805 at the two stations. The main features to note are as follows:

(1) Overall the curves seem to follow the same trends, thus indicating the homogeneity of the record between the two stations;

(2) Between 1715 and 1765 there is a gradual decrease in the number of days of rainfall events. After 1765 there is an increase to a peak around 1785 followed by a fairly rapid decline to the end of the record in 1805;

(3) The most significant feature of the graphs occurs in approximately 1760. Prior to that date the curves of both stations are relatively synchronous, after that date, although the trends remain the same, the curves are clearly separate, with York Factory having distinctly more rainfall days than Churchill.

A thorough examination of the record discounts the possibility that the division of the curves can be attributed to either changes in the observers, or location of the observations. Therefore, it is reasonable to assume that a change in the circulation pattern affecting the two stations occurred resulting in a different rainfall pattern for York Factory. Prior to 1760, both sites experienced the same circulation regime; after that time they were under different regimes.

Before speculating on the causes of this change, it is necessary to show that there is corroborating evidence. If York Factory was in a different climatic regime after 1760, as the rainfall record indicates, there should be changes in other climatic indicators. Figure 2 gives the number of days on which heavy or continuous rainfall was recorded. This category includes rainfall that lasted for more than six hours - usually as a result of cyclonic systems - and intense heavy rain derived from cumulonimbus and heavy cumulus clouds. The curves show that from 1765 to 1800 York Factory experienced more rainfall events of these types than Churchill.

Figure 3 is a plot of the number of days on which thunder and lightning were recorded at the stations. This graph clearly indicates that York Factory had a higher frequency than Churchill from 1757 onwards. Since thunder and lightning are only associated with

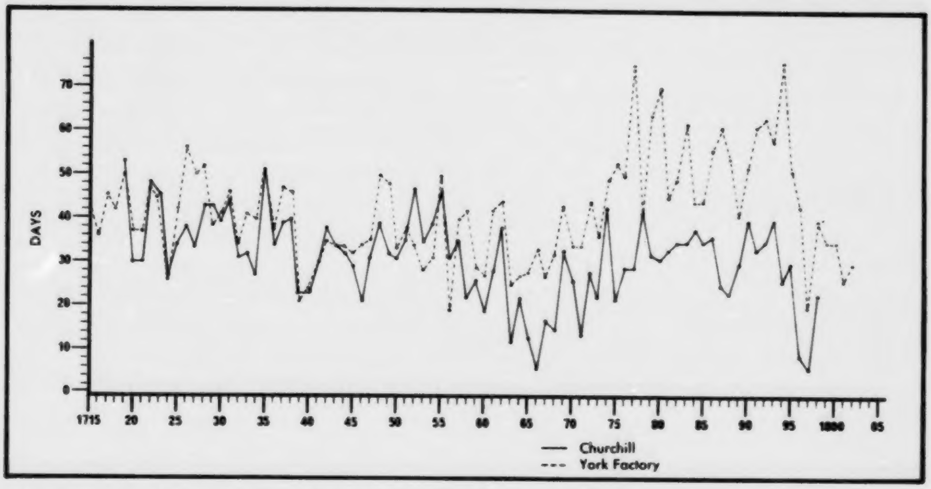


FIGURE 1: *Number of days of rain at Churchill and York Factory (January to December, 1715-1805).*

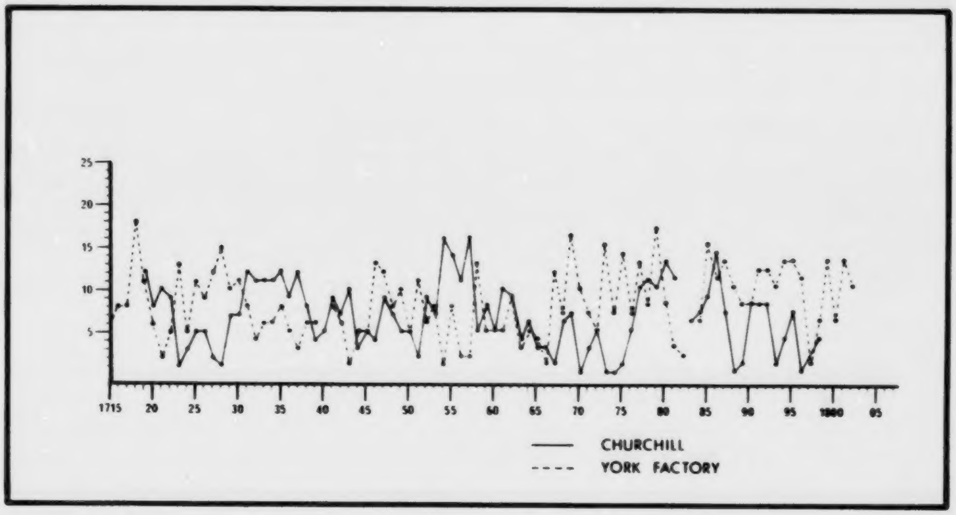


FIGURE 2: *Number of days with heavy or continuous rainfall at Churchill and York Factory (January to December, 1715-1805).*

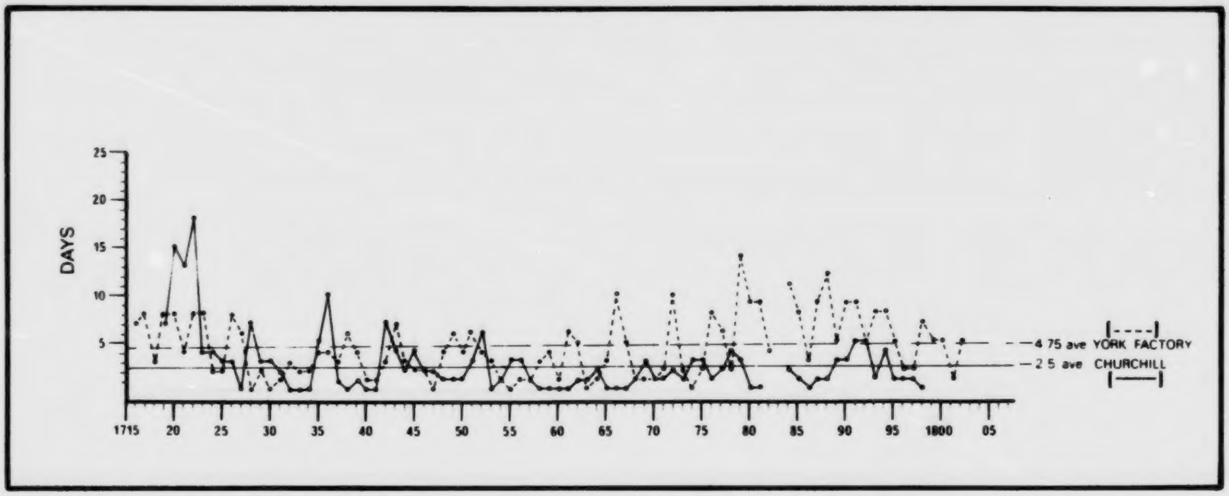


FIGURE 3: *Number of days of thunder at Churchill and York Factory (January to December, 1715-1805).*

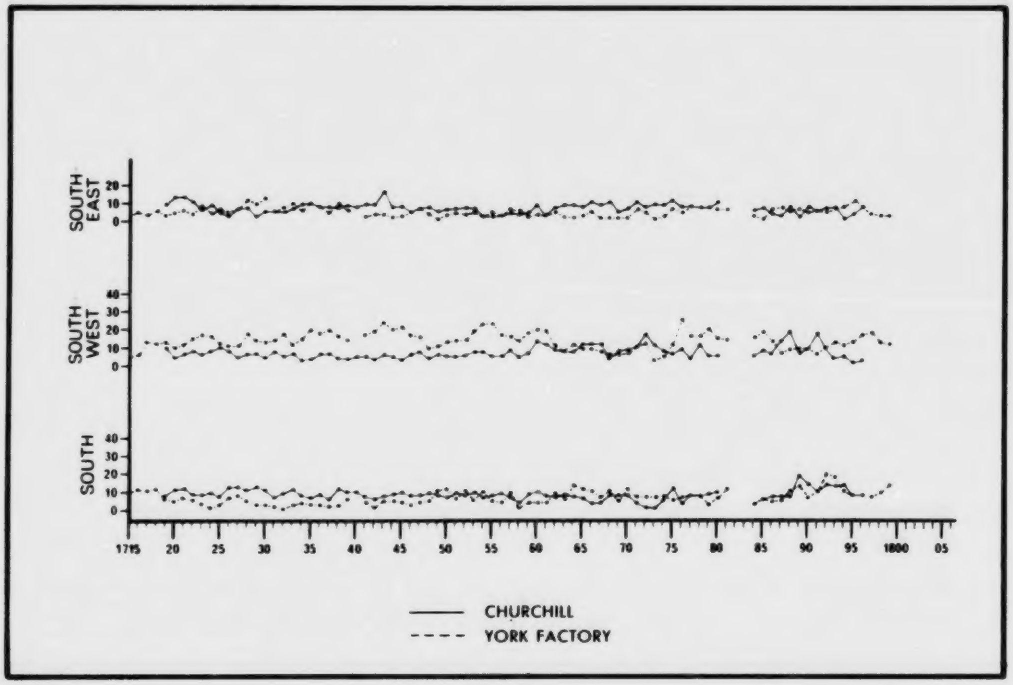


FIGURE 4: *Percent frequency of south wind at Churchill and York Factory (January to December, 1715-1805).*

cumulonimbus clouds it is safe to conclude that York Factory was experiencing greater instability during the summer months after that date.

The most obvious measure of a change in the general circulation is a shift in wind direction pattern at a particular location. Figure 4 shows the percentage frequency of southerly winds at the two stations and Figure 5 the northerly component winds. The original observers recorded the wind direction to 32 points of the compass, but these were reduced to eight according to the method recommended by Conrad and Pollak (1950).

Southerly component winds account for less than 25% of the total wind frequency at both stations therefore changes in any of the three directions are relatively small, nonetheless the following changes can be seen;

(1) There is a noticeable change in the frequency of southerly winds in the period between 1760 and 1765;

(2) There is an increase in the variability of southerly winds at both stations after 1760.

The northerly component winds (Figure 5) have similar changes in percentage frequency. This is particularly noticeable in the northwest and north directions, with the change again occurring around 1760. Because the northerly winds are predominant, accounting for approximately 60% of the annual frequency, the changes they manifest indicate a shift in general circulation.

The similarity of climatic conditions at York Factory and Churchill Factory in the period from 1715 to 1760 suggests that they were under similar climatic conditions. After 1760 the conditions at York Factory change, while those at Churchill remain essentially the same. The changes at York Factory include a shift in the wind-direction patterns with a more meridional component of increased southerly and northerly winds. These are accompanied by an increase in the number of days of rain events, particularly convective rainfall and the associated thunder and lightning.

Apparently, prior to 1760 the mean summer position of the Arctic Front was to the south of York Factory, therefore both stations experienced a tundra-type climate. After 1760 the mean summer position shifted north of York Factory and that station experienced boreal forest conditions, but Churchill remained in the Tundra regime.

There has been much debate about when the Little Ice Age ended, with estimates ranging from 1750 A.D. to as late as 1850 A.D. Obviously different regions, with their unique local conditions that can delay or advance the onset of climate changes, experienced the effects

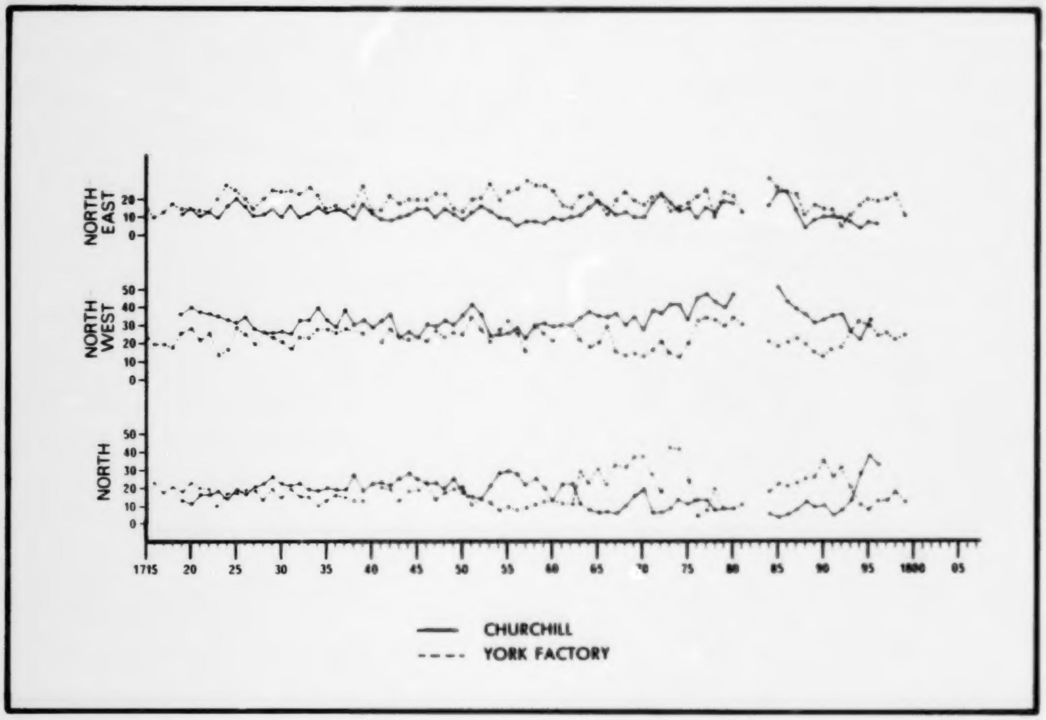


FIGURE 5: *Percent frequency of north wind at Churchill and York Factory (January to December, 1715-1805).*

of the hemispheric warming at different times and with different intensities. The author defers to the general position taken by Lamb that "It is reasonable to regard the time from 1550 to 1700 as the main phase for most parts of the world..." (Lamb 1977, p.463). My data suggest that a change occurred in the general circulation, apparently reflecting a shift in the mean summer position of the Arctic Front. Whether this marks the end of the Little Ice Age is not clear because there are no temperature records to indicate an amelioration. However, it does suggest that there was a modification of temperatures in association with an increase in the southerly flow of air. Certainly there was an increase in rainfall events and therefore in the total amounts of rainfall, and they coincide with similar patterns noted throughout the northern hemisphere and reflected in an increase in the extent of glaciation.

SUMMARY

Meteorological Journals and Daily Journals maintained by the Hudson's Bay Company at York Factory, Churchill Factory and Fort Prince of Wales beginning in 1714 A.D. have provided detailed climatic information. A coding method was developed that allows these historical weather references to be stored in a computer. Analysis of the data includes frequency counts of the number of days of rain events, snow events, thunder and lightning, wind, cloud cover, and frost. Graphs of these frequencies indicate that York Factory and Churchill Factory were experiencing very similar conditions up to 1760. After that year, homogeneity exists between the two stations with regard to the general trend, but there are distinct differences in the frequencies. This change is most clearly seen in the frequency counts of rain events, with York Factory having a greater number of days of rain than Churchill Factory. It is supported by evidence of more days with thunder and lightning at York Factory and a distinct change in the frequency of north and south winds.

It would appear, that, prior to 1760, the mean position of the Arctic Front was south of York Factory, thus placing both stations in the Tundra zone. After 1760, the mean position of the Arctic Front apparently lay between the stations - thus Churchill remains in the Tundra zone while York Factory is now clearly in the Boreal zone.

REFERENCES

Alt, B.T. 1978. Synoptic climate controls of mass balance variations on Devon Island Ice Cap. Arctic and Alpine Research 10:61-80

Barry, R.G. 1967. Seasonal location of the Arctic Front over North America. Geographical Bulletin 9(2):79-95.

Barry, R.G., R.S. Bradley, and J.D. Jacobs. 1975. Synoptic climatological studies of the Baffin Island area. In: Climate of the Arctic. Edited by: G. Weller and S.A. Bowling. Harper and Rowe, New York. pp.82-90.

Brinkmann, W.A.R., and R.G. Barry. 1972. Paleoclimatological aspects of the synoptic climatology of Keewatin, Northwest Territories, Canada. Palaeogeography, Palaeoclimatology, Palaeoecology 11:77-91.

Bryson, R. 1966. Air masses, streamlines and the boreal forest. Geographical Bulletin 9:228-269.

Catchpole, A.J.W., and T.F. Ball. 1981. Analysis of historical evidence of climate change in western and northern Canada. In: Climatic Change in Canada 2. Edited by: C.R. Harington. Syllogeus No. 33:48-96.

Conrad, V., and L.W. Pollack. 1950. Methods in climatology. Harvard University Press, Cambridge. 459 pp.

Dey, B. 1973. Synoptic climatological aspects of summer dry spells in the Canadian Prairies. Unpublished Ph.D. thesis, University of Saskatchewan, Saskatoon. 180 pp.

Fletcher, R.J., and G.S. Young. 1976. Climate of Arctic Canada in maps. Boreal Institute for Northern Studies, Edmonton, Occasional Publication Number 13:1-85.

Lamb, H.H. 1977. Climate, present, past and future. Volume 2. Methuen, London. 835 pp.

Locke, C.W., and W.W. Locke. 1977. Little Ice Age snowcover extent and paleoglaciation thresholds. North Central Baffin Island, N.W.T. Canada. Arctic and Alpine Research 9: 291-300.

Minns. R. 1970. An air mass climatology of Canada during the nineteenth century: an analysis of the weather records of certain Hudson's Bay Company forts. Unpublished Masters thesis, University of British Columbia, Vancouver. 73 pp.

Mitchell, V. 1973. A theoretical tree line in central Canada. Annals of the Association of American Geographers 63(3):296-301.

A RECONSTRUCTION OF NEW ENGLAND CLIMATE USING HISTORICAL MATERIALS, 1620-1980

William R. Baron[1] and Geoffrey A. Gordon[2]

INTRODUCTION

The purpose of our research is to reconstruct climate parameters for the northeastern United States during the historical period (1600-1980). To reconstruct these records we utilize a variety of historical materials including instrumental records, diaries, newspapers, agricultural journals, maritime logs and government documents. We have developed a series of methodological procedures to aid in the compilation, storage, retrieval, synthesis and analysis of these diverse data.

As a result of our research, a number of long-term record reconstructions and chronologies have been assembled. Among these are instrumental records of temperature and precipitation, content analysis indices of temperature and precipitation, chronologies of the frequency of thunder storms, frequency of snowfall, the percentage of days with fair sky and the length of the growing season. The focus of the present study is on southern New England and especially eastern Massachusetts because this was the area settled earliest, and the oldest materials containing climatic information describe conditions pertaining to that area.

SOURCE MATERIALS

To date, no seventeenth century instrumental records have been discovered for the northeast. The only source of climatic information are the quantitative descriptions of months, seasons and years that appear in diaries and letters. For eastern Massachusetts, 58 diaries with usable climate materials were located and analyzed. Typical of these diaries is one kept by John Hull of Boston from 1653-1682 (1857). Hull was an early mint master, treasurer and colonial administrator who described the progress of the seasons, cold spells,

---

[1] Institute for Quaternary Studies, University of Maine, Orono, Maine 04469, and Department of History, Northern Arizona University, Flagstaff, Arizona 86011

[2] Institute for Quaternary Studies, University of Maine, Orono, Maine 04469

storms, floods, droughts, snow storms and their impact on colonial society. The Hull diary is also typical in that it was published, as were a number of other very early diaries used in our reconstruction of the first century of European settlement.

Diaries are available for the eighteenth century, but many are only accessible in manuscript form. On the positive side, daily diaries become more common from the 1730s forward. The more common appearance of daily diaries at that time is perhaps attributable to an increasing population, a more readily available supply of paper and improved economic conditions. More people had the leisure time and financial resources to keep diaries and to buy instruments. We have located several hundred eighteenth century diaries with climatic descriptions. Of these, nearly 100 have daily entries (Baron 1982A). The daily diaries analyzed to reconstruct temperature and precipitation indices are those kept by Benjamin Walker of Boston (1726-1749) and Aaron Wight of Medway, Massachusetts (1777-1819). Both diaries are available only in manuscript form at the Massachusetts Historical Society, Boston. Daily as well as non-daily diaries were used to reconstruct the growing season, snowfall, thunder storm and cloud cover chronologies.

Instrumental records also become available in New England for the first time during the eighteenth century. The Royal Society of London provided the stimulus to undertake keeping of the first instrumental records for New England. A few years of record were kept by Thomas Robbie and Isaac Greenwood, both tutors at Harvard College, during the 1720s and 1730s. However, the honour of keeping the first usable record of any length belongs to Professor John Winthrop (1742-1779) also of Harvard College. He began recording temperature, barometric pressure and rainfall accumulation in the 1740s and continued, with few interruptions in the daily routine, until his death in 1779 (Stearns 1970). For length and accuracy it is probably the best New England instrumental record for the century. Winthrop's record has been reconstructed as part of a larger record compilation for the east coast of the United States (Landsberg *et al.* 1968). Other long records for the latter part of the century include those kept by E. Stiles for Newport and New Haven (1763-1795) and E.A. Holyoke for Salem (1754-1829).

From 1800 through the 1840s instrumental records become more common. These early instrumental records were kept by interested amateurs such as Holyoke for Salem and S. Rodman (1905) for New Bedford, 1812-1860, or by private organizations such as the Rumford Foundation (1818-1839), which kept records for Boston. These data were often published in the popular journals and scientific periodicals of the time. Unfortunately no adequate index of these periodicals exists, so each must be systematically searched to locate all published records. Federally-sponsored record-keeping was limited to that undertaken by the United States Army Surgeon General's Office (1843-1854). It ordered that temperature, and eventually precipitation records be kept at all military posts. While useful, these records lack the detail and consistency of later federal records. Some states, such as New York, supported the establishment of meteorological networks, but for New England only scattered records compiled by the State Agricultural Boards survive from the early nineteenth century. These data are difficult to locate and use in any systematic way, but provide some information.

There are far more records available for the latter half of the nineteenth century. They were kept by a variety of agencies. Sources of data include the *Reports* of the United States Patent Office (1850-1858) and its successor in record keeping, the United States Department of Agriculture (1863-1890); the *Contributions* of the Smithsonian Institution (Schott 1876, 1881); and the *Monthly Weather Review* published by the United States Signal Service (1870-1891) and then by the United States Weather Bureau (1892 forward). In addition, it became more common for agencies of the individual states to keep records. These include *Reports* of the Massachusetts Board of Agriculture (1880-1899). Academics, such as A. Caswell at Brown University in Providence (1882), also kept records, and a regional network sponsored by the New England Meteorological Society was organized in the early 1880s and remained in operation until the late 1890s. The manuscript materials on which many of the federal reports are based is available on microfilm from the United States National Archives (1818-1892). Records kept by states, individual researchers and regional organizations often were published and are available at state archives and libraries. While the existence of many of these records can be documented, a great many were destroyed by fire, or otherwise lost, and are unavailable. Fortunately, some compilations of those data survive in published form (Kirk 1939, Goodnough 1915, 1930).

To complete our reconstructions and chronologies, we used various United States Weather Bureau/NOAA[1] publications to obtain twentieth-century data. The most useful publication was the *Climatological Data Report for the New England Section* that has been published monthly since 1896.

---

[1] United States National Oceanic and Atmospheric Administration.

RECORD RECONSTRUCTIONS

The reconstructions for temperature, precipitation, and frequency of thunderstorms, snowfall, cloud cover, and frost that follow, collectively cover the period from 1621 through 1981. Reconstruction of each different type of record has its own unique pitfalls and problems, and specific methodologies appropriate to the data were developed.

Instrumental Records

Reconstructing records from diverse instrumental materials that are homogeneous and compatible with United States Weather Bureau records for the twentieth century presented a number of methodological problems. The most frequent of these were:

(1) daily mean temperatures were calculated following various methods, some based on three daily readings, taken at varying times, and others on daily maximum and minimum temperatures;

(2) instruments were dissimilar, and there was usually no effort to maintain a standard calibration;

(3) observers, observation locations, and exposures changed frequently;

(4) background on how observations were made, the type of exposures or who made the observations were often unavailable; and

(5) in many instances data were missing or incomplete.

Figure 1 illustrates some of these problems. The Winthrop diary reconstruction (Figure 1A) is based on a record kept by an individual for a period of over 30 years (Winthrop 1742-1779). Unlike later records, observations were taken from a thermometer that was placed indoors in an unheated room on the north side of Winthrop's residence. There are also gaps in the record whenever he went on trips, especially in the middle 1770s when the Boston area was occupied by the British during the American Revolution. In addition, observation times were not standardized, and often several daily readings were taken, but at different times each day. All of these problems increase the difficulty of reconstructing a homogeneous record that will be directly comparable to twentieth-century records for the Boston area. Most of these problems have been resolved for this record (Landsberg *et al.* 1968).

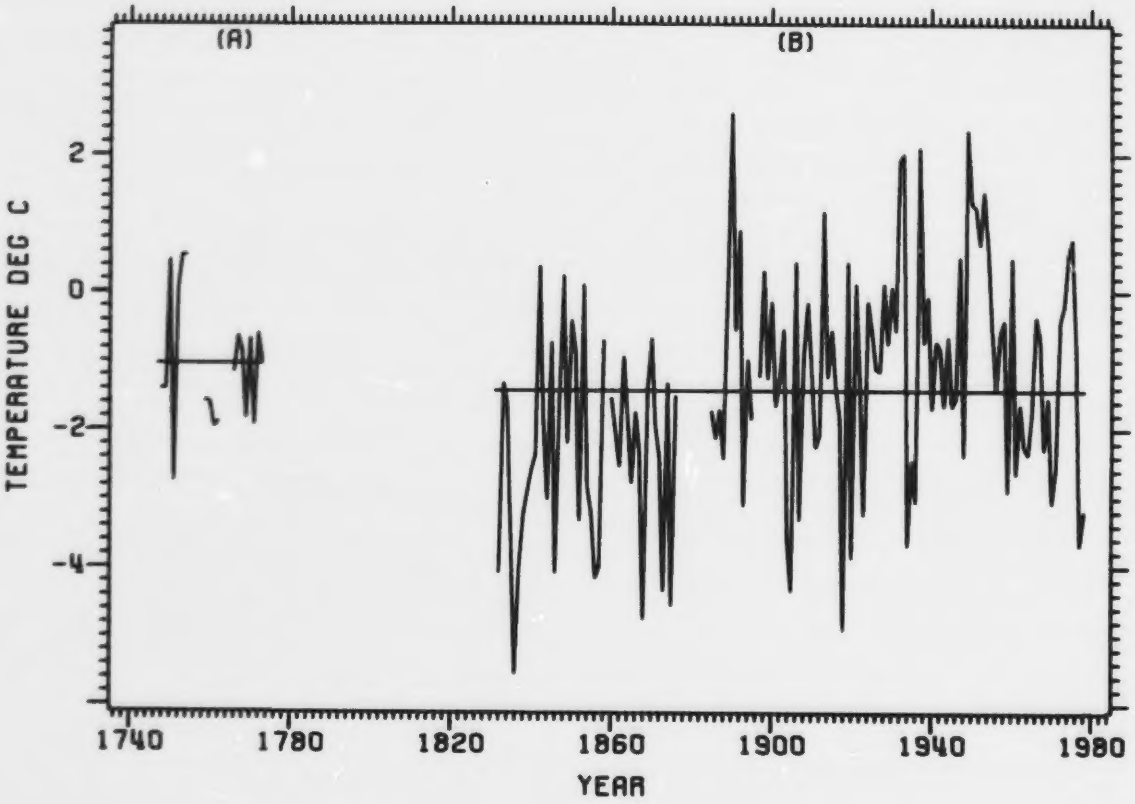


FIGURE 1: *Reconstructed winter (December - February) temperature for Boston (A), and Providence (B). Information is dated by the year of the January month.*

Figure 1B is a compilation of instrumental records for Providence, Rhode Island. This record was maintained by several individuals and agencies. From 1831 to 1876, Caswell (1882) kept the record at Brown University. Readings were taken three times daily at sunrise, 1 p.m. and 10 p.m. The observation site was outside in an open area on a hill overlooking the city. From 1877 through 1884, a precipitation record is available but no temperature record has yet been located. From 1884 through 1906, three separate records were kept by the City Engineer (1884-1905), by Hoyt (1887-1893) and by Upton at the Ladd Observatory (1891-1896). All of these have been published by the New England Meteorological Society. For this period of overlap, the City Engineer's record was selected as being most representative, with daily mean temperatures based on maximum and minimum temperatures taken from self-registering thermometers. The observation site was at a small reservoir and pumping station on a hill overlooking the city. From 1905 forward, records were kept by the United States Weather Bureau. Their instruments were located, first on the roof of a building in the city (1905-1935) and then at garden level beside the building (1936-1953). In 1949, the Weather Bureau started to move its operations to the city airport approximately three to four miles (4.8 to 6.4 km) southwest of the city. From 1953 to the present, all observations have been taken at this site. The instruments are located at ground level, and self-registering thermometers are used to obtain daily maximum and minimum temperatures. Although Figure 1 shows this record as a complete reconstruction, it is not. Rather, it is a compilation of several records for one location. The records have not been homogenized and adjusted to the most recent twentieth century record, but there do not appear to be any major problems with the record. Examples of other instrumental records can be found in Baron *et al.* (1980).

Content Analysis of Daily Diaries

Few instrumental records prior to 1840 were located, and we have used qualitative descriptions of weather to reconstruct an impression of the New England climate. Content analysis techniques have been developed to produce temperature and precipitation indices from these qualitative materials. Figure 2 shows annual temperature and precipitation indices for the Boston area based on qualitative data from the Walker and Wight diaries. Narrative descriptions of the weather were quantified by constructing verbal scales and then

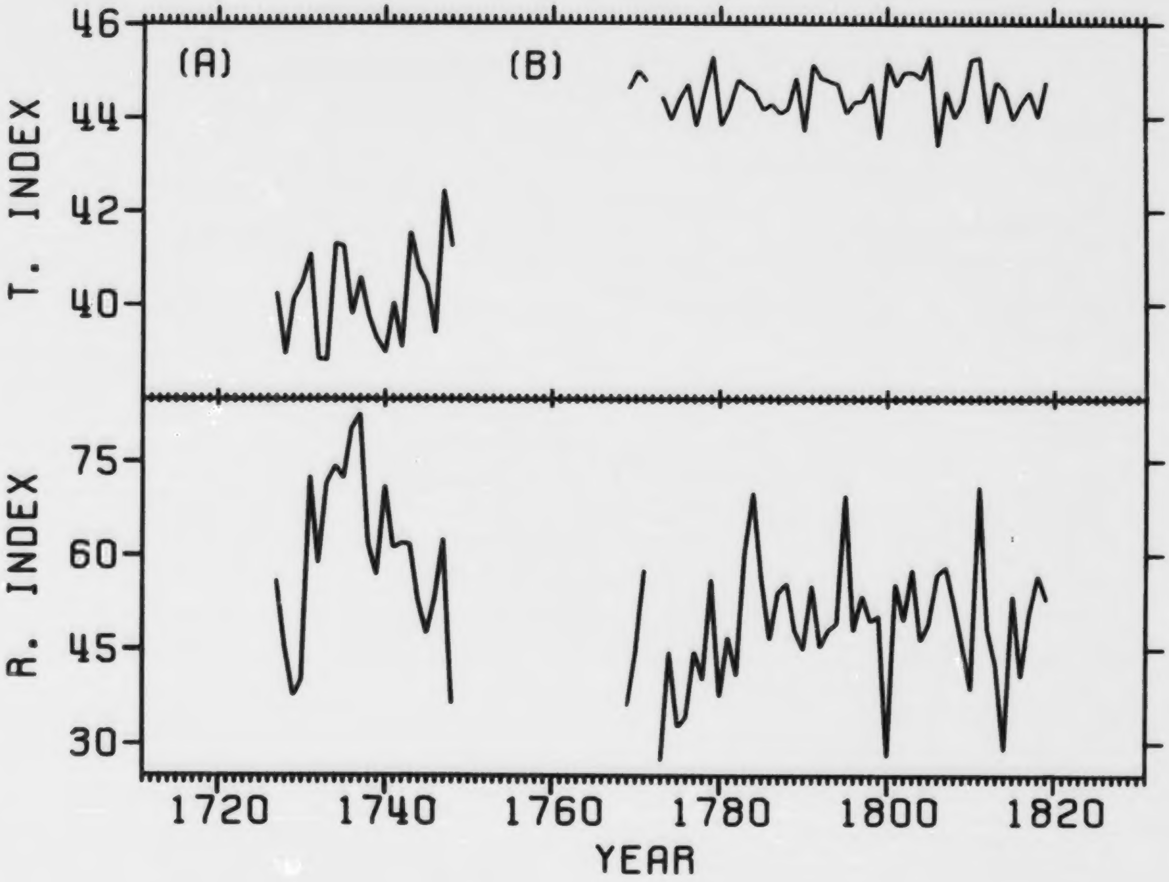


FIGURE 2: *Annual temperature and precipitation indices for the Boston area. Indices were derived from the Walker (A), and Wight (B) daily diaries. The indices are unscaled, see text for details.*

equating these scales to an arbitrary numeric scale. These scales must then be equated among different diarists. Content analysis methodology is discussed in detail in Baron (1980).

A number of factors can influence the results of this type of analysis. The amount of descriptive detail and the frequency of missing entries can affect the accuracy of the indices. In addition, the diarist's physical condition, occupation (whether indoors or out), and age can also influence the perception of temperature and precipitation (Baron 1982A). These factors result in indices that must be rescaled to make them compatible between diaries. This feature is apparent in Figure 2.

Content Analysis of Monthly, Seasonal and Annual Descriptions

Before 1730, daily diaries are not available but many diaries contain qualitative descriptions of the climate for months, seasons and years. In order to quantify this type of narrative information, we modified a methodology developed by Lamb (1965) for use on medieval materials. Word scales were constructed that evaluate temperature and precipitation in terms of departures from normal. A numeric scale was then designed with zero as a mid-point representing normal. The positive side of the index represents warmer or wetter conditions, while the negative side represents cooler or drier conditions. Figure 3 illustrates temperature and precipitation indices for eastern Massachusetts, 1621-1729. Normal conditions may be considered to be represented by the range from +2 to -2. Because there were few individual continuous diaries for this period, descriptions from a number of diaries were grouped together to form a long-term record for a specific region. Some of the major problems inherent in these indices are a noticable decrease in reporting during the autumn, probably due to the pressures of getting in the harvest and, for the 1621-1650 period, a changing perception of what was normal as the earliest settlers became more acquainted with North American climate (Kupperman 1982).

Growing Season Compilation

Many diarists and later meteorological observers recorded dates of the first and last killing frosts. From this information, a chronology describing the parameters of the growing season has been reconstructed (Figure 4). The first part of this reconstruction

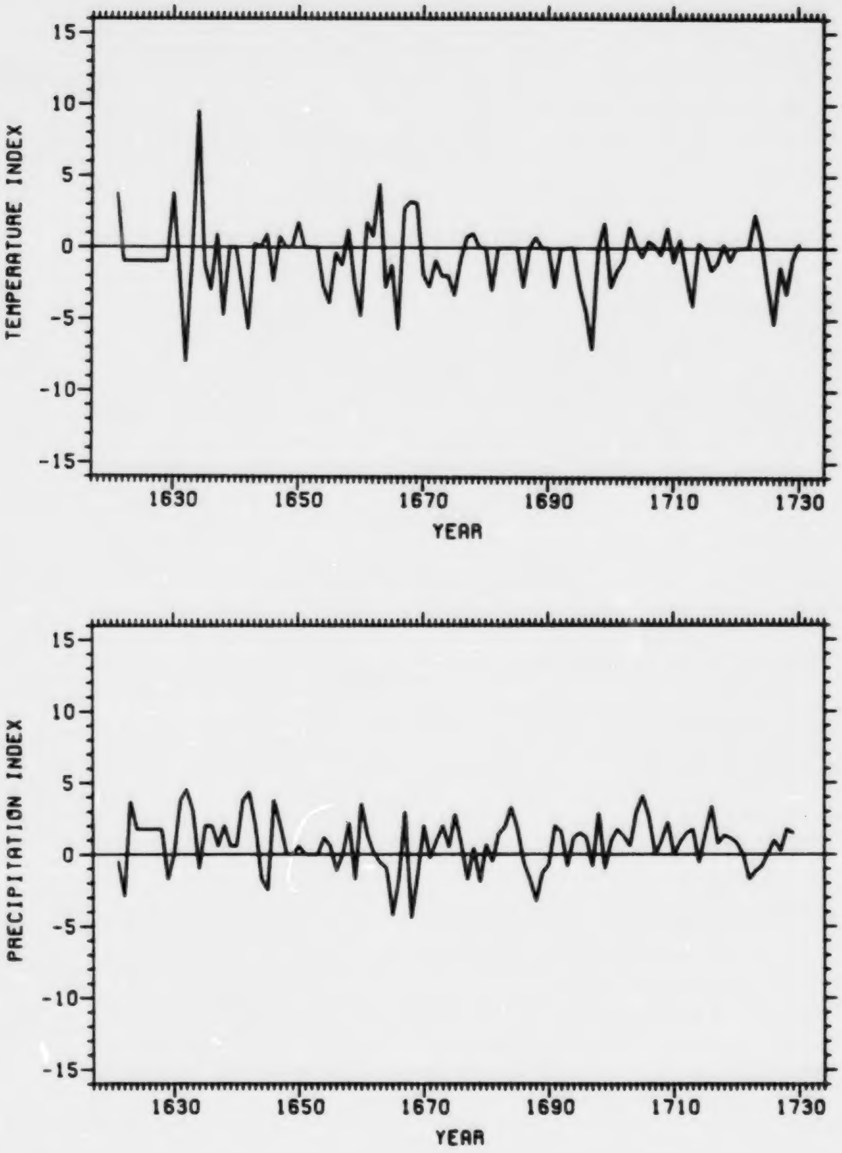


FIGURE 3: *Seventeenth century annual temperature and precipitation indices for eastern Massachusetts. Indices were derived from a variety of non-daily manuscript sources. "Normal" range is considered to be from +2 to -2.*

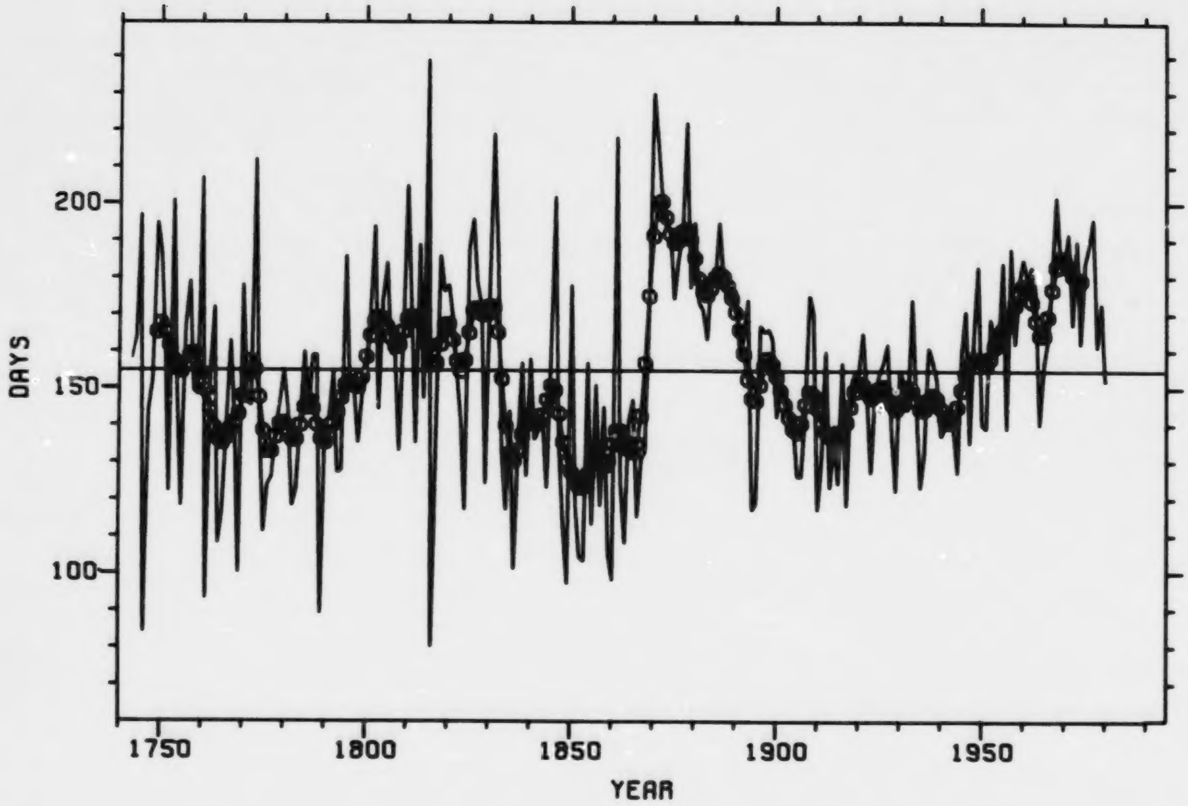


FIGURE 4: *Length of growing season, eastern Massachusetts. Growing season defined as number of days between last and first "killing frost". Open symbols denote low-pass filtered data which passes approximately 50% of the variance at a period of 10 years.*

(to 1840) is based on a large number of diaries from the Boston area at elevations below 200 feet (62 m). From 1840 to 1892, manuscript climatological data from the United States National Archives were used, and after 1892 United States Weather Bureau observations were used. From 1733 to 1947, observations were recorded as "killing frosts" on various agricultural crops such as melons, pumpkins, corn or potatoes. These crops are injured when temperatures fall below 0°C (Chang 1968 referred to in Baron, Gordon *et al.* in press), so that after 1947 a killing frost was reported on the basis of temperatures at or below 0°C. For our compilation for 1948 to 1980, the first and last incidence of 0°C were taken to be comparable to the earlier "killing frost" record.

Cloud-Cover Compilation

A winter cloud-cover compilation for eastern Massachusetts (Figure 5) was reconstructed using daily descriptions of the sky. Data were taken from diaries, agricultural journals, early amateur scientific reports and government records. Observers employed cloud-cover classifications of "fair" (clear), "cloudy" (partly cloudy) and "overcast" (cloudy). The terms in parenthesis were used only in the twentieth century. Each day was classified and tallied as monthly totals and percentages. Monthly data were then converted to a seasonal or annual format and filtered with a low-pass filter. The compilaton is based on very accurate information for the periods 1727 through 1820, 1850 through 1892, and 1896 through 1947. Data from the period 1821 through 1849 is much more scattered making reconstruction of this period very difficult. The United States Weather Bureau changed its reporting format after 1947 to tenths of cloud cover. This new format is not compatible with the rest of the record, so the reconstruction has not been taken beyond 1947. The low-pass filtered data emphasize the fluctuations at time scales greater than 10 years.

Thunderstorm and Snowfall Frequency Compilations

Both of these compilations (Figure 6) for eastern Massachusetts used materials similar to those used for growing season and cloud-cover reconstructions (Baron 1982B), and were filtered. The data were tallied in a monthly and seasonal fashion. In instances where there were several concurrent records, data were grouped and the highest frequency

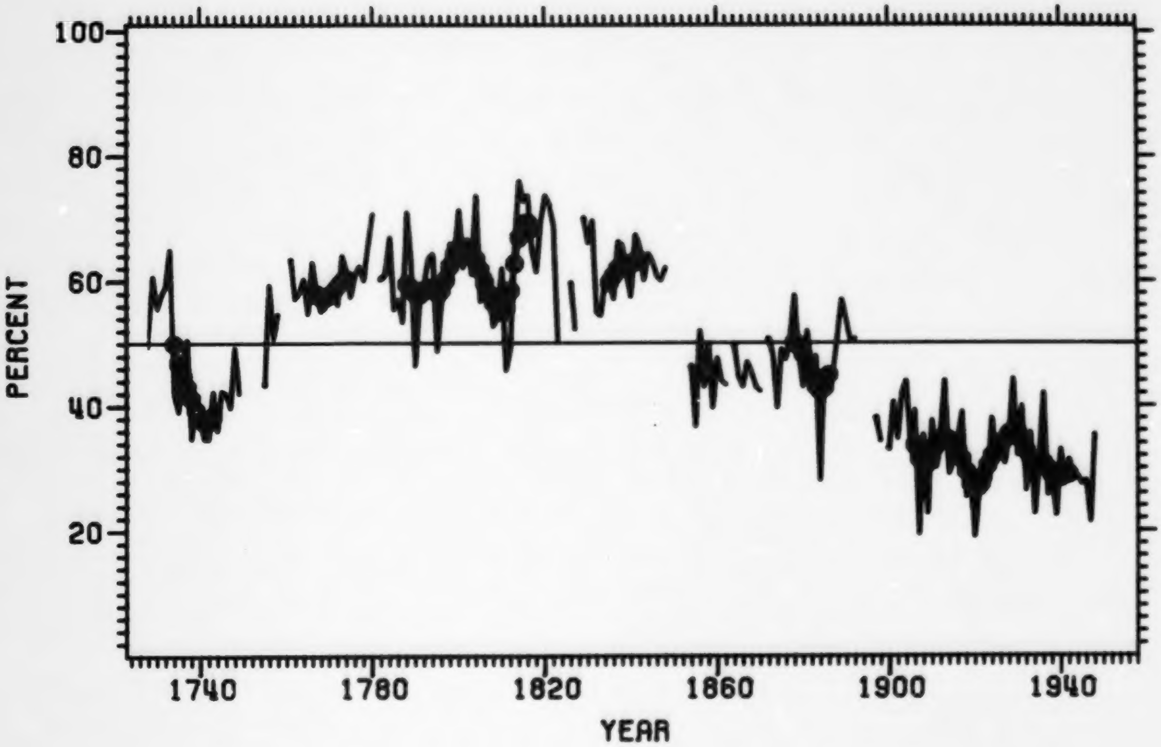


FIGURE 5: *Percentage of winter days with fair-sky conditions for eastern Massachusetts. Open symbols denote low-pass filtered data. See text for details.*

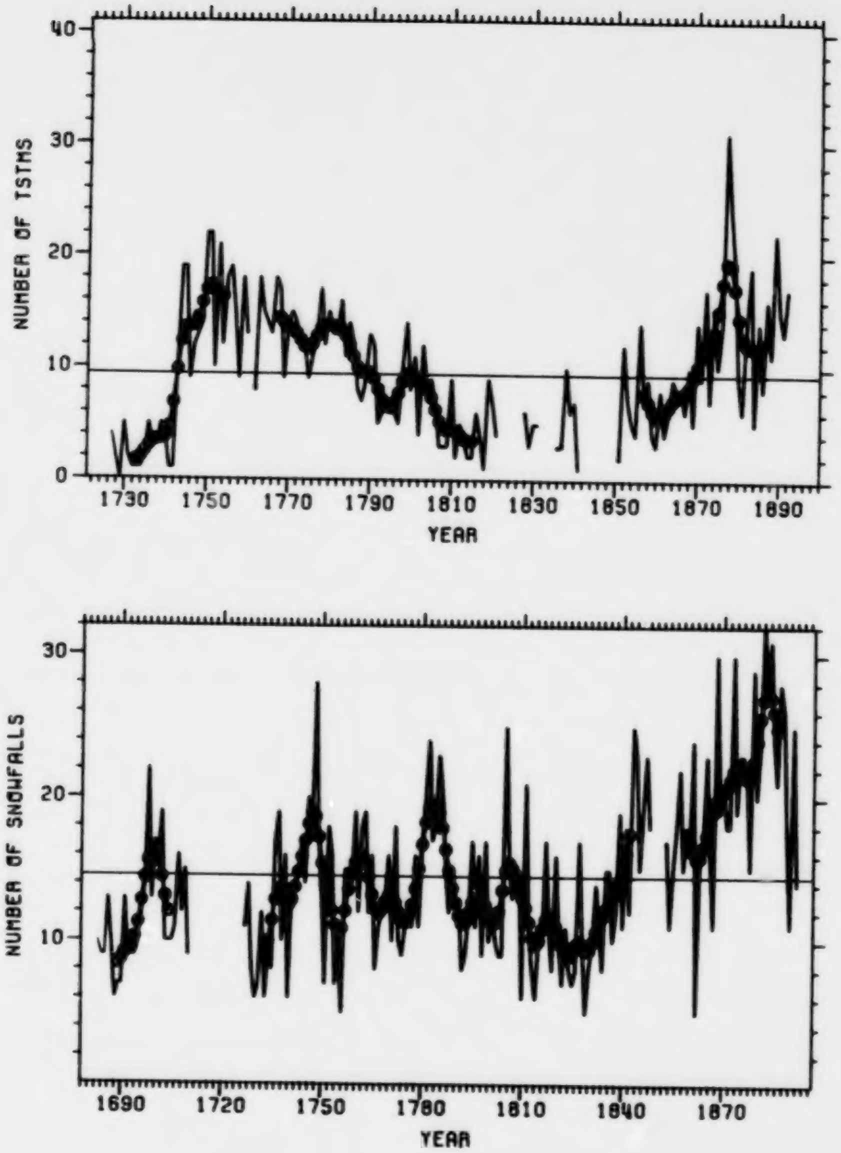


FIGURE 6: *Number of summer (June – August) days with thunderstorms (top), and winter (December – February) days with snowfall (bottom) for eastern Massachusetts. Open symbols denote low-pass filtered data. See text for details.*

recorded. Although some records indicated more than one storm or snowfall, they did not influence the record, as only the days with an event were tallied.

The thunderstorm frequency compilation presented problems. During the early nineteenth century, a number of observers, especially those at military installations, failed to record all thunderstorms. For years when only military records are available (1820-1850), this tendency to under-report may have artificially depressed the frequency curve. The same problem also seems to be evident for the period 1726 through 1745, when only a single diary was available. We are confident of the accuracy of data for all other periods.

ANALYSIS AND CONCLUSIONS

To date, our analysis of these reconstructions has been preliminary and non-statistical. However, we can make some general comments. Our long-run temperature records for New England show evidence of a general warming trend from the 1800s to about 1940. This trend is accompanied, however, by significant variations on a time scale of 10 to 20 years, which in some cases completely overwhelm the longer-term warming. The development of a greater number of long-run instrumental records will permit an examination of the temporal and spatial characteristics of the important short-scale variations. The frost-free season reconstruction shows a decrease in year-to-year variability that apparently accompanied the long-term temperature change. The greater variability of the timing of first and last frost dates prior to about 1860 would have made an impact on the risk factor in New England agriculture. The snowfall reconstruction does not allow comparison with that of the twentieth century, but a significant increase in the frequency of snowfall is apparent in the late 1800s. As with the temperature trends, there are also important 10-year to 20-year variations. Where the trend in temperature is gradual, the increase in snowfall frequency begins much more abruptly at approximately 1830.

The cloud-cover reconstruction reveals a dramatic shift toward fewer days with fair skies. The definition of terms describing cloudiness leave us with an inexact measurement at best, but our evidence indicates that there has been an alteration in the general state of this important climatic factor. The frequency of thunderstorms has undergone significant variations, but not in any simple pattern. Thunderstorm frequency has seen a decline and then a dramatic increase, and there are obvious 10-to 20-year long fluctuations. The apparent coherence between seasons is much stronger for the 10- to 20-year fluctuations in thunderstorm frequency than in snowfall or sky cover.

Collectively, these reconstructions reveal that for eastern Massachusetts the general warming since the mid-1800s has been accompanied by an increased frequency of thunderstorms and snowfall, and an overall tendency to more cloudy conditions. For all the parameters examined, the shorter-term fluctuations are seen to be important to any expectation of conditions in a particular year, and they reveal a persistence in the fluctuation of conditions. It is only through the development of long-term records that these shorter-term fluctuations can be put into perspective.

Future work will involve development of similar records for additional localities in New England. Also, more attention will be given to statistical description of the data - particularly to time-series analysis and an examination of the spatial coherence of the rather dramatic variations seen in the eastern Massachusetts data. Expanding geographic coverage is a particular concern. Development of records for the mid-Atlantic states, the Old Northwest and the Maritime Provinces of Canada will greatly enhance the value of this type of climate reconstruction.

In conclusion, we believe the records reconstructed here will be useful in a variety of ways. The authors would encourage their use by others who are engaged in the reconstruction of past climates. We hope that this type of climatic record will be useful in verifying climatic reconstructions derived from proxy data (e.g. tree-ring and ancient pollen records), for the study of climatic change, and for climate-impact analysis. We seek interaction with and input from other researchers as to the type of climatic reconstructions that are needed and the specific locations that may need further representation.

ACKNOWLEDGEMENTS

We thank the staff of the Northeast Environmental Research Group for their assistance. Research was supported by United States National Science Foundation grants ATM 7908415, ATM 8019514, and ATM 8115714.

REFERENCES

Baron, W.R. 1980. Tempests, freshets and mackerel skies: climatological data from diaries using content analysis. Ph.D. thesis, University of Maine at Orono. 560 pp.

_______. 1982A. The reconstruction of eighteenth century temperature records through the use of content analysis. Climatic Change 4:385-398.

_______. 1982B. Eighteenth century New England climate variation and the suggested impact on society. Maine Historical Society Quarterly 21 (4):201-218.

Baron, W.R., D.C. Smith, H.W. Borns, Jr., J.Fastook, and A.E. Bridges. 1980. Long-time series temperature and precipitation records for Maine, 1808-1978. University of Maine Life Sciences and Agriculture Experiment Station Bulletin 771:1-255.

Baron, W.R., G.A. Gordon, H.W., Borns, Jr., and D.C. Smith. (In press). Frost-free season record reconstruction for eastern Massachusetts, 1733-1980. Journal of Climatology and Applied Meteorology.

Caswell, A. 1882. See station histories for Providence.

Goodnough, X.H. 1915. Rainfall in New England. Journal of the New England Water Works Association 29:237-276.

_______. 1930. Rainfall in New England. Journal of the New England Water Works Association 44:245-306.

Holyoke, E.A. 1754-1829. Manuscript meterological journals for Salem, Maine. Harvard University Libraries, Cambridge, Massachusetts.

Hull, J. 1857. John Hull's Diary. American Antiquarian Society, Transactions and Collections 3:169-249.

Kirk, J.M. 1939. The weather and climate of Connecticut. Hartford, Connecticut, Geological and Natural History Survey, State Geological and Natural History Survey Bulletin 61:1-426.

Kupperman, K.O. 1982. The puzzle of the American climate in the early Colonial Period. American Historical Review 87 (5):1262-1289.

Lamb, H.H. 1965. The Early Medieval Warm Epoch and its sequel. Palaeogeography, Palaeoclimatology, Palaeoecology 1:13-37.

Landsberg, H.E., C.S. Yu, and L. Hang. 1968. Preliminary reconstruction of a long-time series of climatic data for the eastern United States. University of Maryland, Institute for Fluid Dynamics and Applied Mathematics Technical Note BN-571:1-42.

Massachusetts Board of Agriculture. 1880-1899. Reports. Boston, State of Massachusetts. (Issued yearly).

New England Meteorological Society. 1884-1896. Bulletin. Boston, New England Meteorological Society. (Issued monthly).

Rodman, T.R. 1905. Monthly, annual and average temperature and precipitation at New Bedford, Massachusetts, 1813-1904. Climate and Crop Report: New England Section, Annual Summary: 9.

Rumford Foundation. 1818-1839. Manuscript meteorological observations for Boston. American Academy of Arts and Sciences Collection, Harvard University Libraries, Cambridge, Massachusetts.

Schott, C.A. 1876. Tables, distribution, variations of the atmospheric temperature in the United States and some adjacent parts of North America. Smithsonian Contributions to Knowledge 277:1-366.

Schott, C.A. 1881. Tables and results of the precipitation in rain and snow, in the United States, and some stations in adjacent parts of North America and in Central and South America. Smithsonian Contributions to Knowledge 353:1-300.

Stearns, R.P. 1970. Science in the British Colonies of America. University of Illinois Press, Urbana. 465 pp.

Stiles, E. 1763-1795. Manuscript meteorological journals for Newport, Rhode Island and New Haven, Connecticut. Stiles Papers, Yale University Libraries, New Haven.

United States Army Surgeon General's Meteorological Register. 1843-1854. Washington, D.C., United States Government Printing Office. (Issued annually).

United States Department of Agriculture. 1863-1890. Reports. Washington D.C., United States Government Printing Office. (Issued annually).

United States National Archives. 1818-1892. Manuscript nineteenth century climatological records on microfilm T-907.

United States Patent Office. 1850-1858. Reports. Washington, D.C., United States Government Printing Office. (Issued annually).

United States Signal Service. 1870-1891. Monthly Weather Review. Washington, D.C., United States Government Printing Office. (Issued monthly).

United States Weather Bureau/NOAA. 1892+. Monthly Weather Review. Washington, D.C., United States Government Printing Office. (Issued monthly).

_______. 1896+. Climatological Data Report for New England. Washington, D.C., United States Government Printing Office. (Issued monthly with annual report).

Walker, B. 1726-1749. Manuscript diary for Boston. Massachusetts Historical Society, Boston.

Winthrop, J. 1742-1779. Manuscript meteorological records for Cambridge, Massachusetts. Harvard University Library, Cambridge.

Wight, A. 1777-1819. Manuscript diary for Medway. Massachusetts Historical Society, Boston.

PREHISTORIC CULTURAL DISTRIBUTIONS AS AN INDICATOR OF ENVIRONMENTAL CHANGE

J.V. Wright[1]

ABSTRACT

The compilation of nation-wide, archaeological cultural maps[2] for the "Historical Atlas of Canada" has reinforced the position that the major prehistoric cultural constructs across Canada correlate with the major environmental zones. While it has long been recognized that prehistoric people did respond dramatically to significant changes in the environment by expanding or contracting relative to environmental advantages or disadvantages, it has not been as fully appreciated that the same principle may well apply to minor environmental fluctuations. It is argued that archaeological distribution data have an important role to play in the area of paleoenvironmental reconstructions, when it is regarded as a macrofossil roughly equivalent to biological macrofossils. Thus, geographic shifts, through time, of prehistoric cultural entities can be interpreted as responses to environmental changes that may not be detected in the palynological records from the same region. The philosophical debate on the respective roles of environment and culture in the formation of human societies has, in the twentieth century, favoured culture as the dominant factor. Man, however, is subject to natural laws that influence his cultural adaptations. While not advocating simplistic environmental determinism, it is apparent that environment does exert a major influence upon the cultural systems that can be analyzed by archaeology; namely, technology, subsistence, settlement pattern and, to a lesser extent, cosmology.

---

[1] Archaeological Survey of Canada, National Museum of Man, Ottawa, Ontario K1A 0M8

[2] Since the prehistory maps are to appear in 1985 as a section of Volume I of the "Historical Atlas of Canada", it is not possible to publish an article containing the maps in *Syllogeus* prior to releasing of the Atlas.

TREE RINGS

INVESTIGATING THE POSSIBILITY OF A RELATIONSHIP BETWEEN VOLCANIC ERUPTIONS AND TREE GROWTH IN CANADA (1800-1899)

M.L. Parker[1]

INTRODUCTION

The science of dendrochronology did not develop until the twentieth century, and it has been only during the last 20 years that radiographic techniques have been applied to tree-ring analysis to provide ring-density as well as ring-width data. It has been demonstrated that surface temperature correlates better with maximum ring density than it does with ring width for tree-ring series derived from trees growing in the coastal, high elevation and high latitude regions of Canada (Parker 1970, 1976; Parker and Henoch 1971).

Both ring-width and ring-density chronologies provide information about climate in the past. In North America, particularly in western and northern regions, the trees provide proxy climatic records that are often many times longer than instrumental records. Techniques, applications and analytical methods still are being devised and refined, but the ultimate usefulness of dendrochronological research depends on an extensive database of tree-ring chronologies.

Many new ring-width and ring-density chronologies now are becoming available for tree-ring sites located in Canada and the northwestern United States. These chronologies already have proven useful for dating projects and environmental studies (Clague *et al.* 1982; Heger *et al.* 1974; Johnson 1982; Jozsa 1981; Jozsa *et al.* 1979, 1980A, 1980B, 1982; Henoch *et al.* 1975; Parker 1976; Parker *et al.* 1973A, 1974, 1976, 1981, 1982; Parker and Henoch 1971; Parker and Jozsa 1973A, 1973B, 1977; Parker and Kennedy 1973; Stryd 1979).

Most of these studies have used tree-ring data from one or a few sites. This study, however, is designed to use tree-ring data from sites extending from the east side of Hudson Bay to the Pacific Coast and from the Prairies to near the northern tree line. The purpose of the current study is to examine these ring-width and ring-density data to determine if the expected drop in temperature, following the eruptions of Tambora in 1815 and Krakatau in 1883 (Macdonald 1972; Watt 1973; Francis 1976; Lambert 1978; Taylor 1978; Taylor *et al.* 1980; Decker and Decker 1981), is reflected in the tree-ring record.[1]

---

[1] M.L. Parker Company Inc., 606-1150 Burnaby Street, Vancouver, British Columbia V6E 1P2

The ring-width and density data are derived from approximately 135 trees from 15 different sites. Two of the sites have provided samples from two tree species, so there are 17 tree-ring groups. Each "group" includes a ring-width chronology and a maximum-ring-density chronology for a total of 34 separate chronologies. These tree-ring series were averaged into regional and total summaries as described in a later section of this paper. The data were examined for those periods prior to and following the eruptions of Tambora and Krakatau.

No immediately-apparent relationship between tree growth and volcanic eruption was detected on a continent-wide scale. However, the approach of comparing tree-ring data with volcanic eruptions over long periods of time and over extensive geographical areas, seems potentially very useful. Although the tree-ring data may reflect temperature (at least summer temperature) quite well, an examination of the literature illustrates that the relationship between volcanic eruptions and temperature is very complex.

SOME CONSIDERATIONS ABOUT VOLCANIC ERUPTIONS, TEMPERATURE AND TREE RINGS

The role of volcanic activity in shaping the earth's crust (especially as it is related to continental drift), causing climatic changes and influencing the evolution of life on earth has received more attention in recent years (Tazieff 1974; Axelrod 1981). There is wide, but not universal, acceptance that volcanic eruptions have an effect on global surface temperatures (Taylor 1978; Taylor *et al.* 1980). The relationship, in quantitative terms, between volcanic activity and climate is poorly known. The profound consequences of the drastic climatic change that might result from increased volcanic activity, suggest that this relationship needs more attention.

Although records of observations of volcanic activity extend back for hundreds of years in the Old World (Thorarinsson 1970), actual observations of volcanic eruptions in western

---

[1] Ed. note: See Skinner's paper in this volume regarding the impact of the eruption of Krakatau and later major volcanic eruptions on Canadian weather.

North America, for example, are limited. In order to better understand the relationship between factors that effect or record climatic change, there is a need to use proxy data (Hoeller 1982).

This is one of the considerations that led to this study. There are trees growing in Canada and the northwestern United States that exceed 1300 years in age (Parker *et al.* 1984). At least some trees at certain locations record summer temperature. A pronounced reduction in ring width and ring density was observed in white spruce trees growing near Great Whale River (Cri Lake), Québec, for the annual rings put down in 1816 and a few years following (Parker *et al.* 1981). This suggests a reduction in temperature at that time. Confirming evidence is presented by Catchpole and Ball (1981) in their examination of Hudson Bay records of ice conditions in northern rivers and seas.

The year 1816 is the infamous year that has been called "the year without a summer" (Francis 1976). It is the year following the eruption of Tambora in 1815. The logical speculation is that the reduction in tree growth at Cri Lake resulted from lower temperatures that were the result of a dust veil produced by the Tambora eruption. If the dust-veil effect is pronounced and global, it should be reflected in the tree rings of trees growing at other locations in Canada.

Relationship between Temperature and Tree Rings

High correlations have been obtained in comparisons between temperature and maximum ring density of trees growing in Canada - particularly at high elevations and high latitudes. A correlation of $r = 0.79$ resulted from a comparison of August temperature and maximum ring density of Engelmann spruce in Alberta (Parker and Henoch 1971). White spruce maximum ring density matched with August temperature in the Inuvik, Northwest Territories area produced a correlation of $r = 0.75$ (Parker 1976). Maximum ring density of Douglas-fir from Haney, British Columbia gave a high correlation ($r = 0.94$) when compared with May through August average temperature (Heger *et al.* 1974).

Relationship between Temperature and Volcanic Activity

The relationship between temperature and volcanic eruptions may be more difficult to establish than the relationship between temperature and tree-ring growth. There is, of course, a different cause and effect relationship, but there is a more complex set of factors that determine surface temperatures than determine tree growth.

Reference is often made to Benjamin Franklin's observations about the "dry fog" in Europe during the summer of 1783 and his speculation that it was due to volcanic activity in Iceland (Francis 1976; Taylor 1978; Taylor *et al.* 1980). Humphreys (1934) recalls Franklin's account of the "fog" and states that it probably was due to eruptions of Skaptar Jökull in Iceland and Asama in Japan. Francis (1976) notes that "1783, of course, was the year of the eruption of Laki in Iceland, the largest eruption of historic times in terms of the volume of material erupted."

The tree-ring record from sites in northern Canada may confirm the effect of volcanic activity in 1883 on climate in 1894 and several years afterward. The tree ring produced in 1894 by white spruce trees in the Mackenzie Delta has the lowest maximum ring density of all rings in those tree-ring chronologies (Parker 1976).

The effect of the Tambora eruption (1815) also may be confirmed in the tree-ring record of white spruce from Cri Lake, Québec. The ring-width index value for 1816 is the second lowest for the entire series from 1700-1977 (Parker *et al.* 1981), and the ring-width values for 1816, 1817 and 1818 are the three lowest consecutive values for the entire series. The maximum ring density values for 1816 and 1817 are the two lowest consecutive values for the entire series.

Humphreys (1934) states that for one to three years after every volcanic explosion that puts large amounts of dust in the high atmosphere, the average temperature was about 1°C below normal. He calculated that if one cubic kilometre (or less) of rock were properly distributed in the atmosphere, it could reduce the intensity of sunlight by 20%. Taylor *et al.* (1980), in reporting on a study comparing temperature and volcanic eruptions on a global basis, state that a "significant dip in temperature can be found within a few years after the major eruption dates in most of the dust veil temperature superpositions..."

Humphreys (1913) explains the mechanisms behind the cooling effect of volcanic dust. Generally, volcanic dust in the high atmosphere reflects and scatters incoming solar radiation more effectively than it does outgoing terrestrial radiation. This has a net cooling effect.

The evidence suggests that the eruption of Krakatau in 1883 had an effect on global temperature. On this Francis (1976) reports:

> "While most of the world was revelling in the gloriously-hued sunsets that followed the Krakatoa eruption, astronomers at the Montpelier observatory in the south of France were recording a rather more sinister effect. The radiant energy from the sun reaching their recording instrument at ground level dropped by about 20 percent when the pall of dust first arrived over Europe, and their readings remained about 10 per cent below normal for many months."

The eruption of El Chichon in Mexico in March and April of 1982 (Simon 1983) has provided the most recent and well-documented evidence that temperature is affected by volcanic dust. This volcano put 50 times (500 times in localized areas) the normal amount of aerosols into the upper atmosphere, and sunlight reaching different parts of the earth's surface could be reduced by at least 10 to 20%. This has led to predictions by United States National Aeronautics and Space Administration (NASA) experts that the effect of this eruption will be a gradual cooling of between 0.3°C and 0.5°C over the next few years in the northern hemisphere.

Considerations and Complicating Factors

Taylor (1978) states that there is a net cooling of surface temperature by about 1°C for a major eruption. However, an accurate estimate of the exact effect of volcanic activity is complicated by several factors. The size of the dust particles is important: relatively very large or very small particles have a different effect than normal-sized particles. Also important are: chemical composition of extruded material; wind direction; season of eruption; and latitude of the volcano. Humphreys (1937) points out that there even can be a warming (greenhouse) effect from relatively coarse volcanic dust.

It is also important to consider the nature of the tree-ring response. Coniferous tree species used in Canada for tree-ring analysis generally grow only during the summer months, so the time of year of the volcanic eruption is important to consider. Trees of different species and trees growing in different climatic regions respond differently to different climatic factors.

The major problem, however, in this type of analysis, may be in separating the climatic signal due to the dust-veil effect from trends due to other causes and from random noise.

TREE-RING DATA

Method

The methods used to build the tree-ring width and density chronologies have been described (Parker 1967, 1970; Jones and Parker 1970; Parker and Meleskie 1970; Parker and Kennedy 1973; Parker and Jozsa 1973B; Parker *et al.* 1980). X-ray densitometry was used for deriving both ring-width and maximum ring density data. The method used for standardizing and summarizing the tree-ring data is described by Parker *et al.* (1981). The "B & C" type chronology is used. In this form, the growth trend has been removed but the short-term and year-to-year fluctuations are retained.

Tree-ring Sites

The tree-ring sites (Figure 1) extend over a large part of Canada: from Cri Lake, near Great Whale River, Québec on the east side of Hudson Bay to Silver City, Yukon Territory, and as far south as the Olympic Peninsula in Washington State to Prelude Lake in the Northwest Territories. Trees from sites in the boreal forest are white spruce and black spruce. The trees from the West Coast are Douglas-fir and western red cedar, and the trees from the dry interior are Douglas-fir.

Data Format

For this project, tree-ring data for the years 1800 through 1899 were used. The chronologies are ring width and maximum ring density of the "B & C" type (Parker *et al.* 1981). The chronologies are grouped into: (1) site chronologies; (2) regional chronologies; and (3) Canada composites. Data are presented in Appendices[1] in table and bar-graph form for the Canadian composites. The regional chronologies are means of from one to five chronologies. Two tree species were used from the Olympic Peninsula site (Douglas-fir and western red cedar), and white spruce and black spruce were used from the

---

[1] Ed. note: See Parker (1983) for data in Appendices, which are not included here.

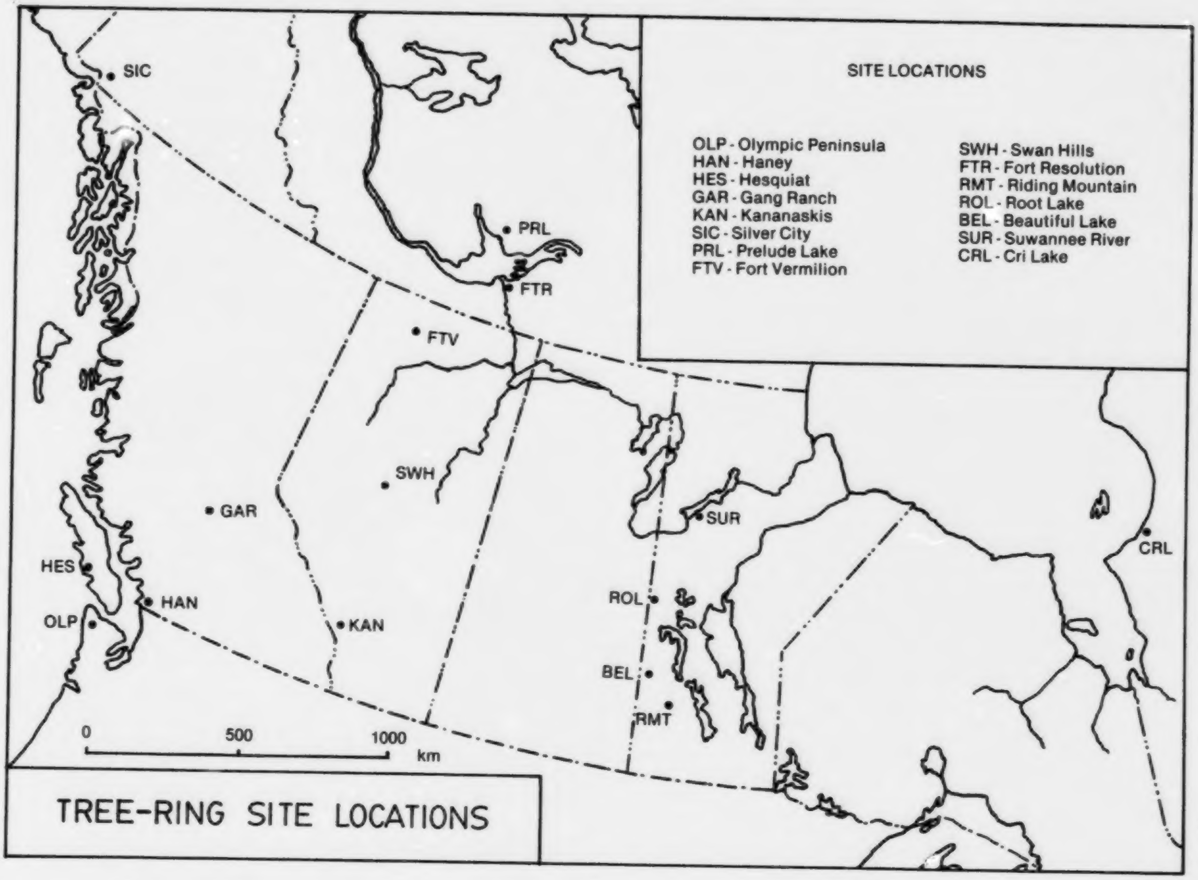

SIC
PRL
FTR
FTV
SWH
GAR
HES
HAN
OLP
KAN
SUR
ROL
BEL
RMT
CRL
0
500
1000
km
SITE LOCATIONS
OLP - Olympic Peninsula
HAN - Haney
HES - Hesquiat
GAR - Gang Ranch
KAN - Kananaskis
SIC - Silver City
PRL - Prelude Lake
FTV - Fort Vermilion
SWH - Swan Hills
FTR - Fort Resolution
RMT - Riding Mountain
ROL - Root Lake
BEL - Beautiful Lake
SUR - Suwannee River
CRL - Cri Lake
TREE-RING SITE LOCATIONS

FIGURE 1: *Tree-ring sites in central and western Canada mentioned in the text.*

Suwannee River site. The Canada Composite consists of mean index values of all 17 chronologies.

Analysis

A visual inspection was made of bar graphs presented in the Appendices. There seem to be patterns of fluctuations of a cyclical nature particularly as shown in the Canada Composite series.

The time and magnitude of the major volcanic eruptions of the nineteenth century are presented in Figures 2 and 3. The dust veil index (DVI) is from Lamb (1970) (taken from Taylor 1978).

If there are effects of volcanic activity on the tree-ring series, and there may well be, these effects are obscured by other influences that form the chronology patterns. There are periods of reduced ring width and density for periods other than those following the major eruptions.

The periods just before and just after the eruptions of Tambora (1815) and Krakatau (1883) are examined. Figure 4 shows, in bar-graph form, the ring-width and maximum ring density composite for: (A) the three years before the Tambora eruption; (B) the year of the Tambora eruption; and (C) the three years after the Tambora eruption. Data for the regional and summary chronologies are presented.

Data show that there is no well-defined, continent-wide reduction in ring width and density for the periods after the eruptions of Tambora and Krakatau.

CONCLUSIONS AND RECOMMENDATIONS

There is ample evidence that volcanoes can be very destructive in terms of human lives and property (Furneaux 1965; Crandell 1980; Macdonald 1972, 1975; Lambert 1978; Thorarinsson 1979). For this reason, it is important to have information about the frequency, duration, magnitude and kind of volcanic activity that has occurred in the past. Of more significance to people in the long run, however, may be the effect that volcanoes can have on global climate.

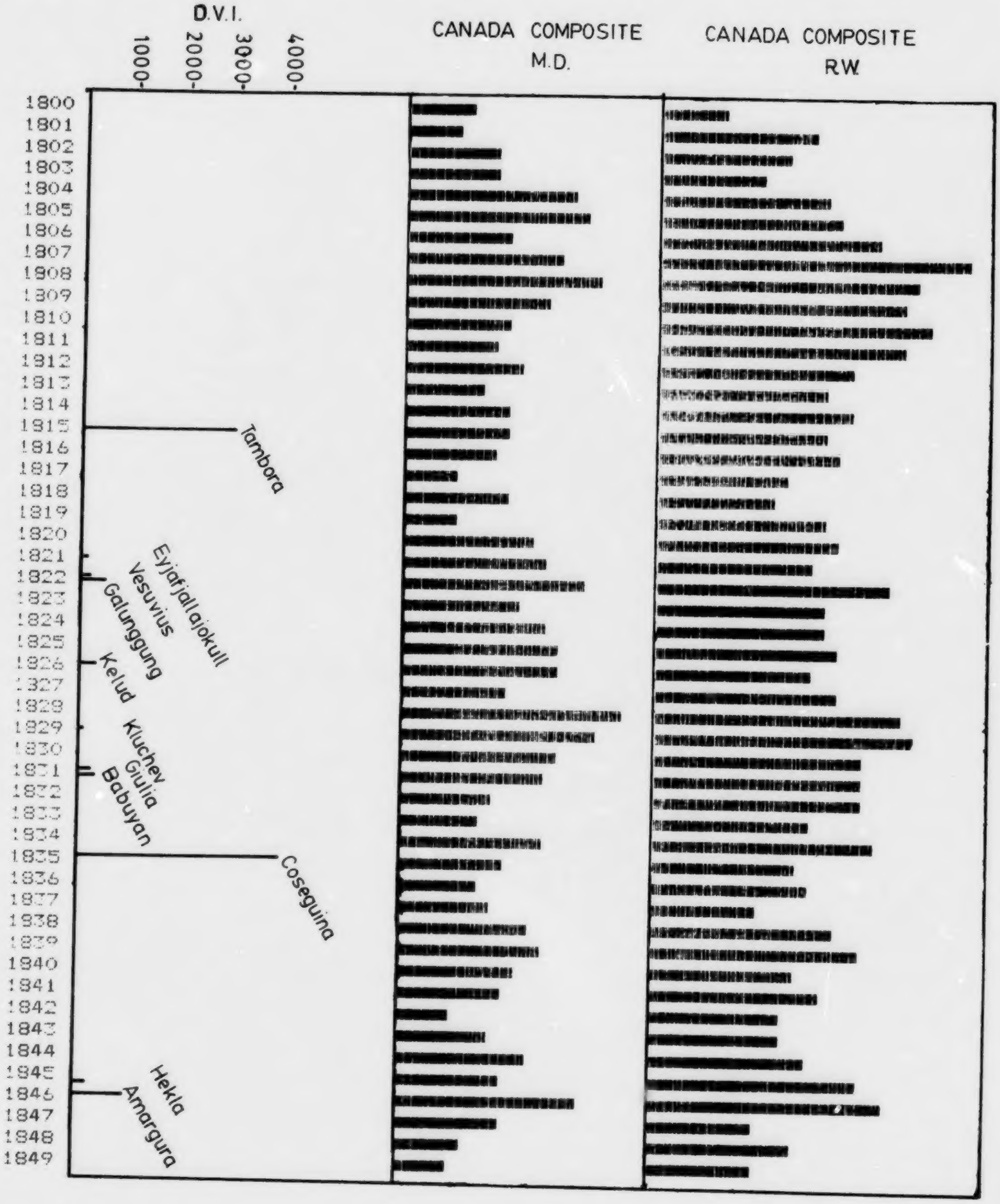


FIGURE 2: *Bar graphs of Canada Composite ring width (RW) and Canada Composite maximum ring density (MD) compared with dust veil indices (DVI) of major volcanic eruptions of the nineteenth century (1800-1849).*

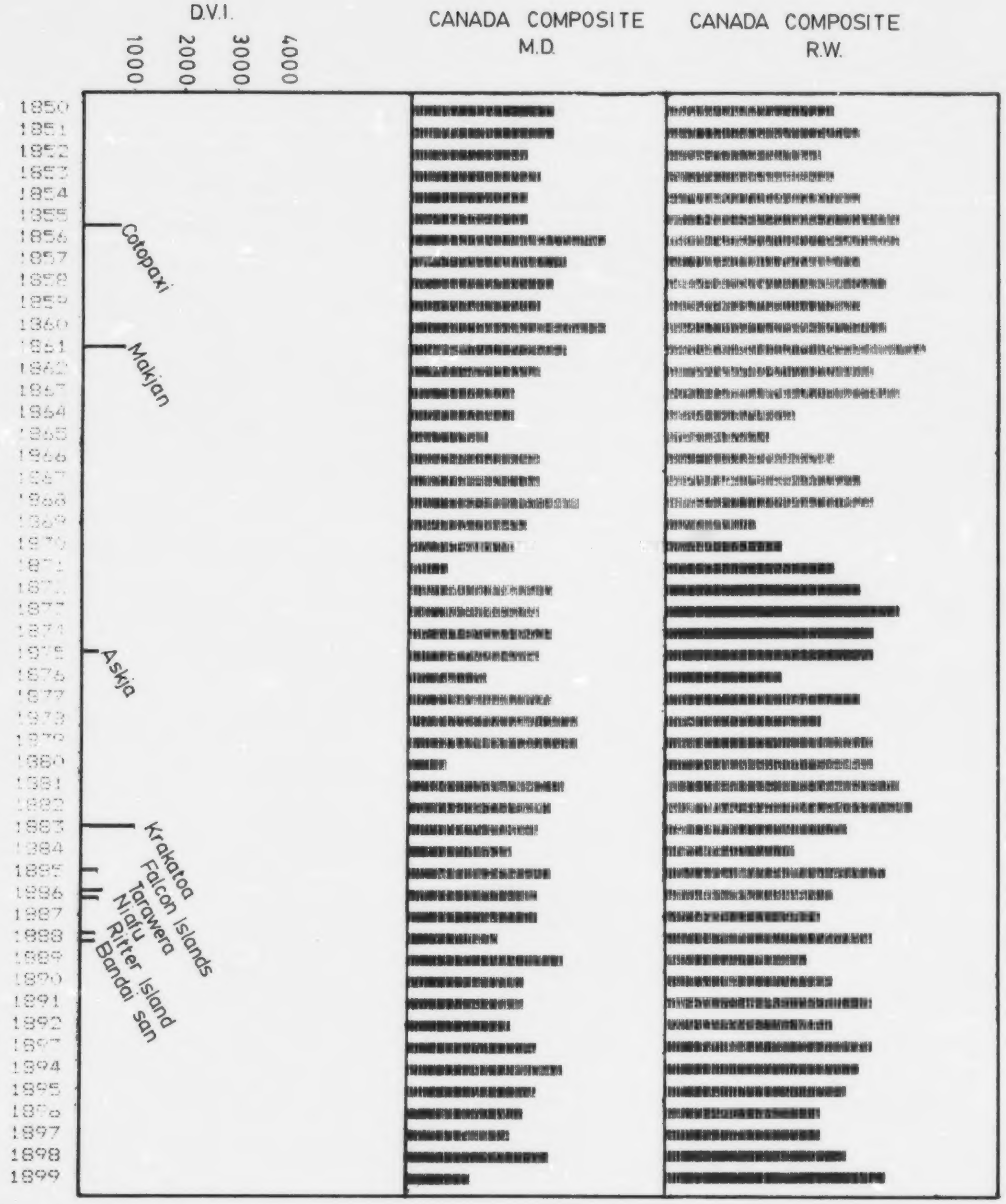


FIGURE 3: *Bar graphs of Canada Composite ring width (RW) and Canada Composite maximum ring density (MD) compared with dust veil indices (DVI) of major volcanic eruptions of the nineteenth century (1850-1899).*

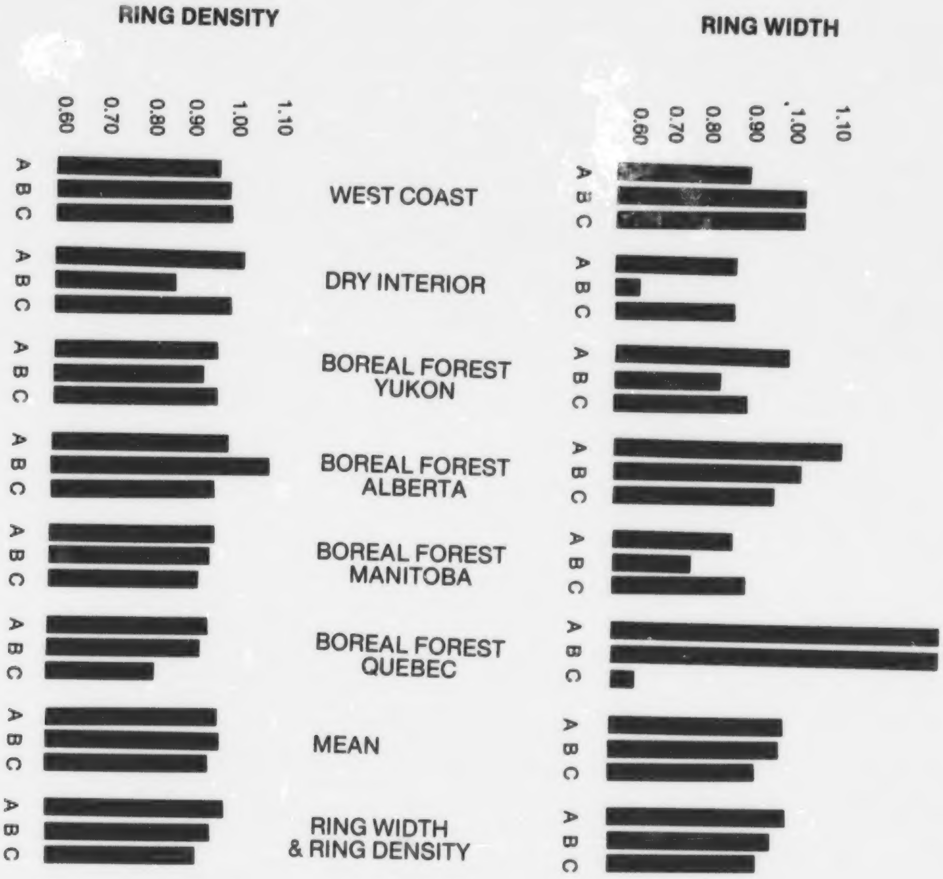


FIGURE 4: *Bar graphs showing means of: (A) the three years before Tambora erupted; (B) the year Tambora erupted (1815); and (C) the three years after Tambora erupted. Data are for six regional summaries, for the Canada Composite series (MEAN) and for the Canada Composite for both ring width and maximum ring density (RW, MD).*

If, as stated by Humphreys (1934), one cubic kilometre of rock material in the atmosphere could reduce temperatures by 20%, what would be the effect of volcanoes that produce 100 times that amount of extrusive material?

Axelrod (1981) states that climatic changes caused by volcanic activity may have led to important extinctions of plant and animal species during past geological periods and led to major changes in species that survived. There is a relationship, in time, between volcanic activity and significant biological changes of life on earth.

Layser (1980) describes major effects that could result from a mean global surface temperature change as small as 1°F (0.5°C). This would cause, for example, great changes in the growing season and migration of the limits of the boreal forest, as well as changes in forest production. The cooling effect might be accentuated by higher albedo (more reflected solar radiation) due to increased snow cover. According to Harris (1983), we can expect the volcanoes in the Pacific Northwest to continue to erupt in the future. Volcanic activity has taken place with greater violence during the last few thousand years than during historic times.

One approach to the study of past volcanic activity and its relationship with climate is through the use of proxy data. Tree rings, ice cores, deep-sea sediment cores, pollen, soils, historic records and archaeological evidence can all yield proxy climatic data (Hoeller 1982). All are interrelated through their common relationship to climate.

I recommend that there should be a continuing effort to compile proxy data of all types and to compare these types with one another. Although tree-rings can yield one of the best forms of proxy data, it is apparent from the results of this study that the relationship between tree growth and volcanic activity is difficult to measure because of the many factors involved.

How can this relationship can be further investigated? In addition to comparing tree-ring indices with volcanic events, dendrochronology can be used in other ways to reconstruct volcanic history. Smiley (1958) used tree-ring analysis to date the volcanic eruption that formed Sunset Crater in Arizona. I have examined well-preserved wood specimens that were covered by volcanic ash from different locations in California, British Columbia, and the Yukon Territory. Dendrochronology has been used to date driftwood deposited on the shore of a lake formed by glacial movement in the Yukon Territory (Clague *et al.* 1982); and tree-ring analysis was used to date driftwood from climatically-related raised beaches in the Hudson Bay area (Parker *et al.* 1982). The eruption of Mount

St. Helens in 1980 provided ample evidence of how vast quantities of wood can be buried by mud and ash following a volcanic eruption (The Daily News...1980). Stumps of white spruce trees (Parker *et al.* 1973) up to 3000 years old are buried in sediment in the Mackenzie Delta. Perhaps the flood sediment layers around the stumps may contain identifiable volcanic ash.

The tree-ring record can be related to volcanic activity in a number of ways and this record could be extended back in time for several thousand years. More refined records of the dust-veil effect need to be devised in order to have good-quality data to compare with the tree-ring record. A broader base of tree-ring site chronologies is needed, and refined statistical techniques are required to separate the dust-veil effect from other factors recorded in tree-ring chronologies.

SUMMARY

Tree-ring data from approximately 135 trees from 15 sites extending from the east side of Hudson Bay to the Canadian Pacific Coast were combined to form various annual ring width and maximum ring density chronologies. These tree-ring summary series, extending from 1800 to 1899, were examined to determine if the climatic response to such eruptions as Tambora (1815) and Krakatau (1883) are reflected in tree growth. No immediately-apparent relationship was detected on a continent-wide scale. However, the volcanic dust-veil effect may be obscured by other more influential causal factors. A possible response is suggested for trees from sites near the tree line in northern Quebec and northern Manitoba.

ACKNOWLEDGEMENTS

I appreciate the assistance of Mal Berry, Joan Masterton, and Bruce Findlay of the Canadian Climate Centre, Atmospheric Environment Service, in organizing this project. I also thank Brian Billings, Public Archives, and Paul Bramhall, Forintek Canada Corporation, for providing me with the tree-ring data tables. This project was financed by the Canadian Climate Centre and the Department of Supply and Services.

REFERENCES

Axelrod, Daniel I. 1981. Role of volcanism in climate and evolution. Geological Society of America, Special Paper 185:1-59.

Catchpole, A.J.W., and T.F. Ball. 1981. Analysis of historical evidence of climatic change in western and northern Canada. In: Climatic Change in Canada 2. Edited by: C.R. Harington. Syllogeus No. 33:48-96.

Clague, John, L.A. Jozsa, and M.L. Parker. 1982. Dendrochronological dating of glacier-dammed lakes; an example from Yukon Territory, Canada. Arctic and Alpine Research 14 (4):301-310.

Crandell, Dwight R. 1980. Recent eruptive history of Mount Hood, Oregon, and potential hazards from future eruptions. United States Department of the Interior, Geological Survey Bulletin 1492:1-81.

Decker, Robert, and Barbara Decker. 1981. Volcanoes. W.H. Freeman and Company, San Francisco. 244 pp.

Furneaux, Rupert. 1965. Krakatoa. Secker & Warburg, London. 244 pp.

Francis, Peter. 1976. Volcanoes. Pelican Books, Harmondsworth. 368 pp.

Harris, Stephen. 1983. In the shadow of the mountains. Pacific Northwest 17 (1):24-33.

Heger, L., M.L. Parker, and R.W. Kennedy. 1974. X-ray densitometry: a technique and an example of application. Wood Science 7 (2):140-148.

Henoch, W.E.S., D.N. Outhet, and M.L. Parker. 1975. Some ice-induced landforms in the Mackenzie Delta. In: Further Hydrologic Studies in the Mackenzie Valley. Environmental - Social Committee, Northern Oil Development, Report No. 74-35:81-110.

Hoeller, A.E. 1982. The role of environmental and historical evidence in climate reconstruction: a preliminary review and appraisal. Atmospheric Environment Service, Canadian Climate Centre, Report No. 82-3:1-113.

Humphreys, W.J. 1913. Volcanic dust and other factors in the production of climatic changes, and their possible relation to ice ages. Bulletin of the Mount Weather Observatory 6 (Part 1), No. 511:1-34.

_______. 1934. Volcanic dust in relation to climate. Transactions, American Geophysical Union, Reports and Papers, Volcanology. pp. 243-245.

Johnson, L. Ward. 1982. Tree rings that tell stories. British Columbia Lumberman, January 1982:1-2.

Jones, F.W., and M.L. Parker. 1970. Geological Survey of Canada tree-ring scanning densitometer and data acquisition system. Tree-Ring Bulletin 30 (1-4):23-31.

Jozsa, L.A. 1981. Dating a landslide in the Mt. Cayley area by analysis of tree-rings. Contract report to the Geological Survey of Canada. Forintek Canada Corporation. 5 pp.

Jozsa, L.A., M.L. Parker, and P.A. Bramhall. 1980A. Feasibility of tree-ring dating at Ozette. Contract report (80-543) for the Washington Archaeological Research Center. Forintek Canada Corporation. 22 pp.

Jozsa, L.A., M.L. Parker, P.A. Bramhall, and S.G. Johnson. 1982. Impact of climatic variation on boreal forest biomass through the use of tree-ring analysis. ENFOR Project No. P-149. 47 pp.

Jozsa, L.A., M.L. Parker, P.A. Bramhall, R.M. Kellogg, and S. Rowe. 1979. Wood and charcoal sample analysis for the Kitwanga National Historic Site. In: Kitwanga Fort National Historic Site, Skeena River, British Columbia. Historic Research and Analysis of Structural Remains. Edited by: George F. MacDonald. Parks Canada, Manuscript Report No. 341:191-211.

______. 1980B. Wood and charcoal sample analysis for the Kitwanga National Historic Site. Contract report (WR 156-79) for Parks Canada, Western Region. Forintek Canada Corporation. 22 pp.

Lamb, H.H. 1970. Volcanic dust in the atmosphere; with its chronology and assessment of its meteorological significance. Philosophical Transactions of the Royal Society (London) 266:425-533.

Lambert, M.B. 1978. Volcanoes. Douglas & McIntyre, North Vancouver.

Layser, Earle F. 19[illegible]0. Forestry and climatic change. Journal of Forestry (November 1980): 678-682.

Macdonald, Gordon A. 1972. Volcanoes. Prentice-Hall, Englewood Cliffs. 510 pp.

______. 1975. Hazards from volcanoes. In: Geological Hazards. Edited by: B.A. Bolt, W.L. Horn, G.A. Macdonald and R.F. Scott. Springer-Verlag, New York.

Parker, Marion L. 1967. Dendrochronology of Point of Pines. M.A. thesis, University of Arizona. 168 pp.

______. 1970. Dendrochronological techniques used by the Geological Survey of Canada. In: Tree-Ring Analysis with Special Reference to Northwest America. Edited by: J. Harry G. Smith and John Worrall. University of British Columbia, Faculty of Forestry, Bulletin No. 7:55-66.

______. 1976. Improving tree-ring dating in northern Canada by X-ray densitometry. SYESIS No. 9:163-172.

______. 1983. Investigating the possibility of relationship between volcanic eruptions and tree growth. Contract report to Canadian Climate Centre, Downsview. 123 pp.

Parker, M.L., Paul A. Bramhall, and Sandra G. Johnson. 1982. Tree-ring dating of driftwood from raised beaches on the Hudson Bay Coast. In: Climatic Change in Canada 3. Edited by: C.R. Harington. Syllogeus No. 49:220-272.

Parker, M.L., R.D. Bruce, and L.A. Jozsa. 1980. X-ray densitometry at the WFPL. Forintek Canada Corporation, Technical Report No. 10:1-18.

Parker, M.L., H.W.F. Bunce, and J.H.G. Smith. 1974. The use of X-ray densitometry to measure the effects of air pollution on tree growth near Kitimat, British Columbia. Proceedings of the International Conference on Air Pollution and Forestry, Marianske Lazne, Czechoslovakia. 15 pp.

Parker, M.L., and W.E.S. Henoch. 1971. The use of Engelmann spruce latewood density for dendrochronological purposes. Canadian Journal of Forest Research 1 (2):90-98.

Parker, M.L., K. Hunt, W.G. Warren, and R.W. Kennedy. 1976. Effect of thinning and fertilization on intra-ring characteristics and Kraft pulp yield of Douglas-fir. Applied Polymer Symposium No. 28:1075-1086.

Parker, M.L., and L.A. Jozsa. 1973A. Dendrochronological investigations along the Mackenzie, Liard and South Nahanni Rivers, Northwest Territories - Part I: Using tree damage to date landslides, ice-jamming, and flooding. In: Hydrologic Aspects of Northern Pipeline Development. Environmental-Social Committee, Northern Pipelines, Task Force on Northern Oil Development, Report No. 73-3, Technical Report 10:313-464.

______. 1973B. X-ray scanning machine for tree-ring width and density analysis. Wood and Fiber 5 (3):192-197.

Parker, M.L., and L.A. Jozsa. 1977. What tree-rings tell us. Canadian Forestry Service, Forest Fact Sheet. 4 pp.

Parker, M.L., L.A. Jozsa, and R.D. Bruce. 1973A. Dendrochronological investigations along the Mackenzie, Liard, and South Nahanni Rivers, Northwest Territories - Part II; Using tree-ring analysis to reconstruct geomorphic and climatic history. Technical report to Glaciology Division, Water Resources Branch, Department of the Environment, under the Environmental-Social Program, Northern Pipelines. 104 pp.

Parker, M.L., L.A. Jozsa, Sandra G. Johnson, and Paul A. Bramhall. 1981. Dendrochronological studies on the coasts of James Bay and Hudson Bay. In: Climatic Change in Canada 2. Edited by: C.R. Harington. Syllogeus No. 33:129-188.

______. 1984. Tree-ring dating in Canada and the northwest U.S. In: Quaternary Dating Methods. Edited by: W.C. Mahaney. Elsevier, Oxford, New York, Tokyo. pp. 211-225.

Parker, M.L., and R.W. Kennedy. 1973. The status of radiation densitometry for measurement of wood specific gravity. Proceedings of International Union of Forest Research Organizations (IUFRO), Division 5 meetings in Cape Town and Pretoria, South Africa, September and October, 1973. 17 pp.

Parker, M.L., and K.R. Meleskie. 1970. Preparation of X-ray negatives of tree-ring specimens for dendrochronological analysis. Tree-Ring Bulletin 30 (1-4):11-22.

Parker, M.L., J. Schoorlemmer, and L.J. Carver. 1973. A computerized scanning densitometer for automatic recording of tree-ring width and density data from X-ray negatives. Wood and Fiber 10 (2):120-130.

Simon, Cheryl. 1982. Red sky at night. Science News 122:120-122.

Smiley, Terah L. 1958. The geology and dating of Sunset Crater, Flagstaff Arizona. In: Guidebook of the Black Mesa Basin. Edited by: R.Y. Anderson and J.W. Harshbarger. New Mexico Geological Society, Socorrow. pp. 186-190.

Stryd, A.H. 1979. The Lillooet archaeological project laboratory analysis program (Part 2). A statement to the Social Sciences and Humanities Research Council of Canada for research grant 575-1241. Cariboo College, Kamloops. pp. 21-24.

Taylor, Billie Louise. 1978. Volcanic eruptions and long-term temperature records: an empirical search for cause and effect. M.A. thesis, University of Toronto. 189 pp.

Taylor, Billie L., Tzvi Gal-Chen, and Stephen H. Schneider. 1980. Volcanic eruptions and long-term temperature records: an empirical search for cause and effect. Quarterly Journal of the Royal Meteorological Society 106:175-199.

Tazieff, Haroun. 1974. The making of the earth, volcanoes and continental drift. Saxon House, Westmead. 111 pp.

The Daily News, Longview, Washington and The Journal-American, Bellevue, Washington. 1980. Volcano, the eruption of Mount St. Helens. Longview Publishing Company, Longview. 96 pp.

Thorarinsson, Sigurdur. 1970. Hekla, a notorious volcano. Almenna Bokafelagid, Reykjavik. 54 pp.

______. 1979. On the damage caused by volcanic eruptions with special reference to tephra and gases. In: Volcanic Activity and Human Biology. Edited by: Payson D. Sheets and Donald K.Grayson. Academic Press, New York. pp. 125-159.

Watt, K.E.F. 1973. Tambora and Krakatoa: volcanoes and the cooling of the world. Saturday Review of Science 4 (January):43-44.

SUMMER DEGREE-DAYS SINCE 1574 IN NORTHWESTERN CANADA AND ALASKA

Gordon C. Jacoby, Jr.[1], Linda D. Ulan[1] and Edward R. Cook[1]

ABSTRACT

Since the 1940s, many researchers have contributed to understanding climatic information recorded by northern trees. Annual increments of tree growth have been related to temperature, precipitation, barometric pressure and ice formation. Degree-days are one of the climatic parameters most closely related to tree growth. Calibrating with degree-days from three stations, tree-ring data have been used to reconstruct degree-days for central Alaska and northwestern Canada since 1574. The reconstruction verifies, and also corresponds with subarctic glacial and ice-core data. A gradual warming trend since about 1840 also has superposed decade-scale trends as well as higher-frequency variations. This and similar reconstructions will afford Global Climate Modellers and radiation modellers the opportunity to examine longer quantitative histories of temperature and insolation.

---

[1] Lamont-Doherty Geological Observatory, Columbia University, Palisades, New York 10964

FOSSIL POLLEN

THE POSTGLACIAL DEVELOPMENT OF VEGETATION IN NEWFOUNDLAND AND EASTERN LABRADOR-UNGAVA: SYNTHESIS AND CLIMATIC IMPLICATIONS

Joyce Brown Macpherson[1]

INTRODUCTION

This paper is an attempt to synthesize inferences on Holocene climatic change based on palynological and related evidence from Canada's eastern seaboard. The evidence has been obtained by many workers and my own field work has been confined to eastern Newfoundland, particularly the Avalon Peninsula in the southeast of the island. This peninsula forms the southeastern extremity of the boreal forest zone of North America, although it lies in the same latitude as Switzerland (Figure 1). To the south, a mixed coniferous-hardwood forest occupies the Maritime Provinces of southeastern mainland Canada and extends into adjacent New England. The boreal forest is dominated by spruce, mainly *Picea mariana*, balsam fir, *Abies balsamea*, and birch, mainly *Betula papyrifera*, with some pine, mainly *Pinus strobus*, on the island. The forest is succeeded northward by a broad zone of boreal woodland or taiga, where open woodland, mainly spruce, is floored with *Cladonia* lichen. Northward again is a belt of forest tundra, a mosaic of open woodland, lichen-heath tundra, and bog, with thickets of alder, willow and dwarfed conifers. Trees decrease in size and frequency northward, and the vegetation of northernmost Labrador is classified as low arctic - a mixture of lichen-heath and sedge-shrub tundra. The latitude here is that of Oslo.

The southward displacement of vegetation zones here compared with their European counterparts is accentuated along the coasts. The cold water of the Labrador Current reduces the temperature of onshore winds, producing frequent low cloud and fog and inhibiting forest growth. Thus the vegetation of much of coastal Labrador and of extreme northern and southern Newfoundland is best described as forest tundra. By contrast, the interior lowland at the head of Lake Melville in the boreal woodland zone of central Labrador enjoys relatively warm sunny summers and supports closed-crown boreal forest.

---

[1] Department of Geography, Memorial University of Newfoundland; St. John's, Newfoundland A1B 3X9

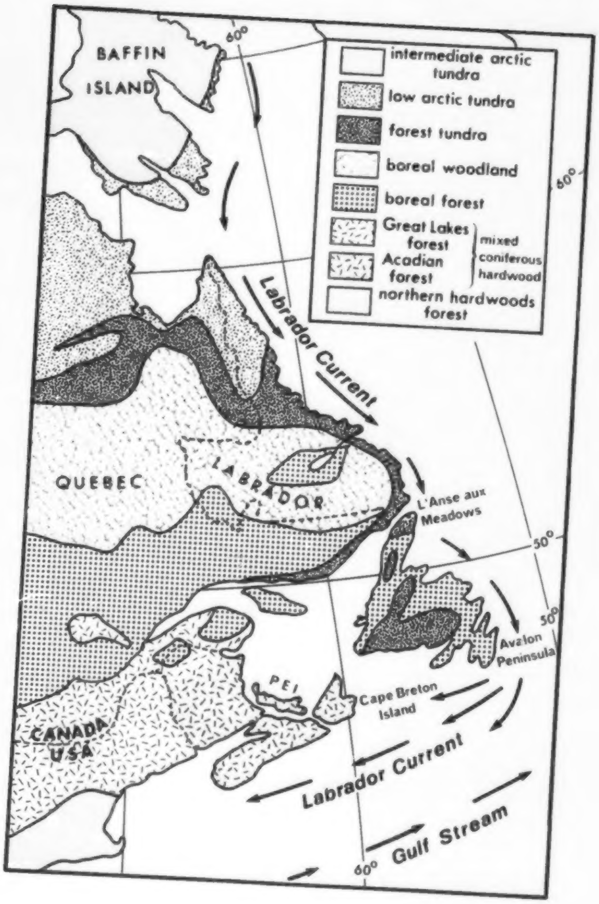

BAFFIN ISLAND
60°
intermediate arctic tundra
low arctic tundra
forest tundra
boreal woodland
boreal forest
Great Lakes forest
Acadian forest
mixed coniferous hardwood
northern hardwoods forest
60°
Labrador Current
QUEBEC
LABRADOR
L'Anse aux Meadows
50°
50°
Avalon Peninsula
PEI
Cape Breton Island
CANADA
USA
Labrador Current
Gulf Stream
60°

FIGURE 1: *Eastern Canada: generalized vegetation regions (after Richard (1980); Rowe (1972); Young (1971)).*

MODERN POLLEN ASSEMBLAGES

Simplified modern pollen assemblages from these vegetation zones, with the exception of the coastal forest tundra, are plotted in latitudinal order in Figure 2. Spectra from lake muds were used in the compilation wherever possible, as the bulk of the dated historical evidence has been obtained from lake sediments. Spectra from Baffin Island to the north are included for comparison; they are dominated by herbs and low shrubs, with some conifer pollen which is transported northward by wind currents. Herb and shrub pollens are still dominant in the low arctic tundra of northern Labrador-Ungava, but birch pollen increases, and toward the boundary with the forest-tundra, so do alder and spruce. Alder, birch and spruce are the dominant pollen taxa of the forest tundra, while boreal woodland spectra may be identified by the very high proportion of spruce. The boreal forest is marked by significant representation of balsam fir pollen; the forest zone on the island of Newfoundland has higher birch percentages than that of Labrador. Spectra from the mixed forest of Cape Breton Island are easily distinguishable by the presence of significant proportions of temperate tree pollen, including maple, beech, ash and oak. No evidence has yet come to light in Newfoundland of fossil pollen spectra similar to those from Cape Breton Island, so it must be assumed that even during the most favourable conditions of the Holocene the vegetation of the island remained boreal in character.

Although these data are presented in a form resembling a conventional pollen diagram it is not with the intention of implying that today's vegetation developed by a simple northward shift of vegetation zones. Species migrated into the region at different times and by different routes, so that a fossil pollen assemblage need not have a modern analogue. Alder, for instance, was expanding in southern Labrador by 8000 yr B.P. (Lamb 1980), while it may not have been present in the Avalon Peninsula before 7000 yr B.P. Balsam fir, on the other hand, was present on the Avalon Peninsula by 8000 yr B.P., although its eastward movement across southern Québec-Labrador occurred between 7000 and 5000 yr B.P. (Jordan 1975; Mott 1976; Lamb 1980). Spruce did not achieve its greatest northward expansion in Labrador until after 4500 yr B.P. A series of maps showing generalized pollen assemblages from sites in Newfoundland and Labrador at selected Holocene time-horizons has been presented elsewhere (Macpherson 1981). The maps demonstrate the existence of fossil pollen assemblages, and hence of the past plant communities, which cannot be matched in the region today.

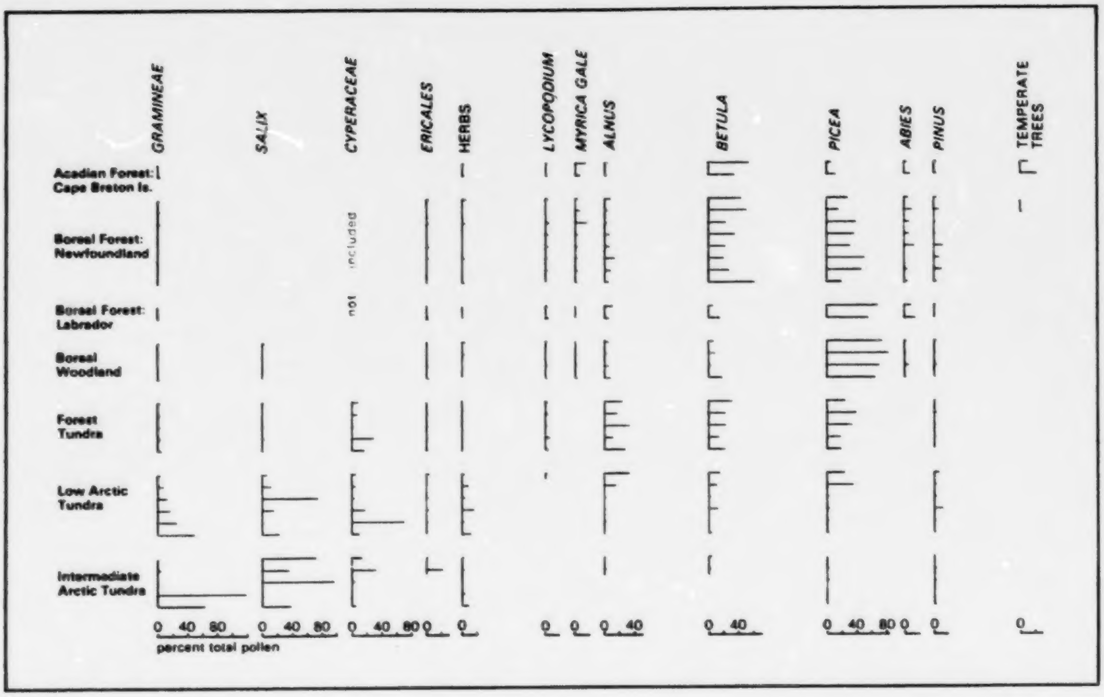


FIGURE 2: *Representative modern pollen spectra, eastern Canada. Data from Andrews* et al. *(1980); Bartley (1967); Jordan (1975); Lamb (1980 and pers. comm.; McAndrews* et *Samson (1977); Mott (1974); Railton (1973); Short and Nichols (1977); Terasmae (1976); and the author.*

HOLOCENE CLIMATE INFERRED FROM FOSSIL POLLEN

Nevertheless, it is possible to draw climatic inferences from pollen and related evidence, as a consideration of Figure 3 will demonstrate. The figure is a simplified percentage pollen diagram from Sugarloaf Pond, a lake at about 100m a.s.l. near the Atlantic coast of the Eastern Avalon Peninsula. The basal organic sediment has been dated at 9270 yr B.P. (GSC - 2601). The stiff laminated clay at the base of the core suggests that sediment began to accumulate in the pond while it was still receiving glacial meltwater. Evidence from other sites on the peninsula indicates that deglaciation proceeded by downwasting, so that tundra-like vegetation was present in the area to colonize the catchment of the lake as it became free of glacial ice, as is evident from the high percentages of tundra-type pollen in the basal sediment. Pollen concentrations were low at first, but by 9300 yr B.P. concentrations increased with the development of a shrub-dominated vegetation. Shrub birch (indicated by grains less than 20 µm in diameter (Ives 1977)), club moss, sweet gale and juniper peaked in succession. After 8500 yr B.P., there was an increase in the proportion of larger birch pollen grains, but most of these fall in the size range which may represent either a shrub or tree origin. No modern analogues exist for these shrub-dominated pollen assemblages, without alder, in terms of either composition or productivity. Summers by 8500 yr B.P. must have been quite warm, perhaps as warm as today's.

By this time, spruce pollen was sufficiently abundant to indicate the local presence of spruce trees, and balsam fir arrived about two centuries later. However, again the vegetation has no modern analogue, despite the presence of the major constituents of today's boreal forest. Some shrub birch was still present, suggesting that the vegetation was open. The presence of *Populus* pollen (probably from trembling aspen), together with charcoal, in almost every sample between 8300-5400 yr B.P. suggests that fire was frequent, and that trembling aspen played the role of a burn colonizer. This tree is most frequent today in north-central Newfoundland, where summers are drier, sunnier and warmer than on the oceanic Avalon Peninsula (Page 1972). Growing-season temperatures of the order of 1°C warmer than present with substantially lower precipitation are indicated for this period. It was not until after 5400 yr B.P. that the pollen assemblage from this site became similar to modern surface spectra from the boreal forest. The increase in balsam fir and cessation of the continuous representation of poplar, accompanied by a reduction in the frequency of charcoal, suggest the establishment of moister conditions than those which prevailed before.

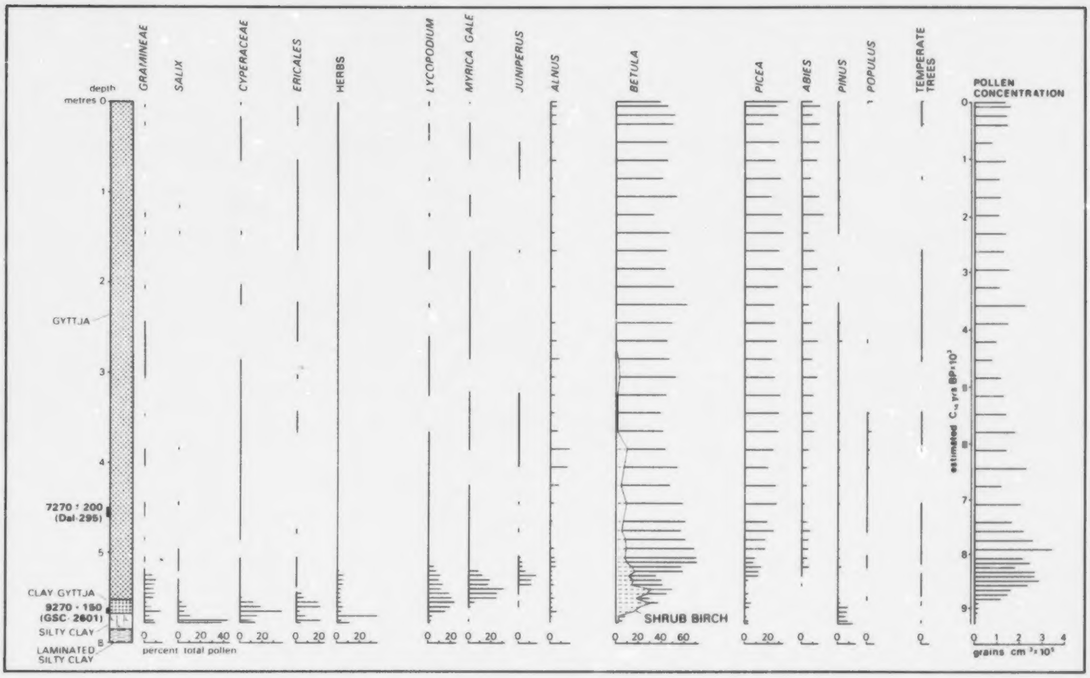


FIGURE 3: *Simplified pollen diagram, Sugar Loaf Pond, Avalon Peninsula, Newfoundland.*

Further evidence of Holocene climatic conditions on the Avalon Peninsula is provided by a site from the present treeline at 220m in the interior upland. The same trends in pollen representation described for the site near St. John's occur here also, except that shrubs and herbs make a greater proportional contribution throughout, and that tree birch declines after 3000 yr B.P. (Macpherson 1982). Rates of pollen influx and lacustrine productivity provide strong evidence of optimal conditions from 5500 yr B.P. to 3000 yr B.P. (Figure 4). Even before the local climatic optimum, conditions were more favourable than those of today, corroborating the summer warmth inferred from the pollen assemblages at the site closer to sea level. Similar evidence reported by Lamb (1980) for a site in the boreal woodland of southern Labrador suggests a warming trend in the early Holocene, with rapid amelioration after 6500 yr B.P. and deterioration at about 2500 yr B.P. Optimal conditions of both warmth and moisture occurred in the tundra of northern Labrador in the same period, 6500-3000 or 2500 yr B.P. (Short and Nichols 1977; Short 1978).

Inferences of variations in temperature and precipitation drawn from similar evidence have been made for much of eastern Canada, but only in relative terms. The sequences of inferred changes are portrayed diagrammatically in Figure 5. The relative nature of the indications of warmth and moisture is emphasized, as is their specificity to each area. Inferred oceanic and atmospheric warm periods are also indicated, and the sources of the information are listed. Temperatures throughout much of the region are more strongly in phase with the oceanic evidence than with the atmospheric. Only in the New England mountains is there evidence of a decline in temperature as early as the end of the atmospheric optimum. Inferences of relative moisture follow two main patterns. In the New England mountains and the Maritime Provinces, conditions following the period of greatest warmth seem to have been moister than those before or during the warmest period. In interior Québec, northern Labrador and Baffin Island, the periods of greatest warmth and greatest moisture coincided. The Avalon Peninsula shows a pattern intermediate between these two. For about 3000 years, between about 8500 and 5500 yr B.P. the climate appears to have been relatively warm and dry. The onset of the period of maximum warmth was accompanied by an increase in precipitation, and moisture has continued to be abundant throughout the cooler period of the last three millenia.

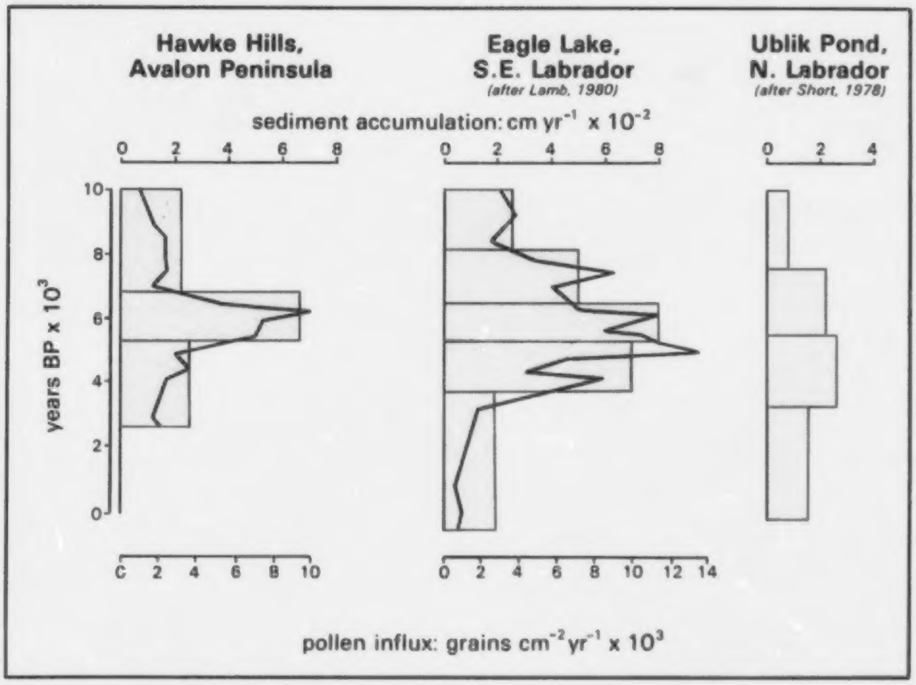


FIGURE 4: *Holocene sedimentation rates and pollen influx for selected lakes in Newfoundland.*

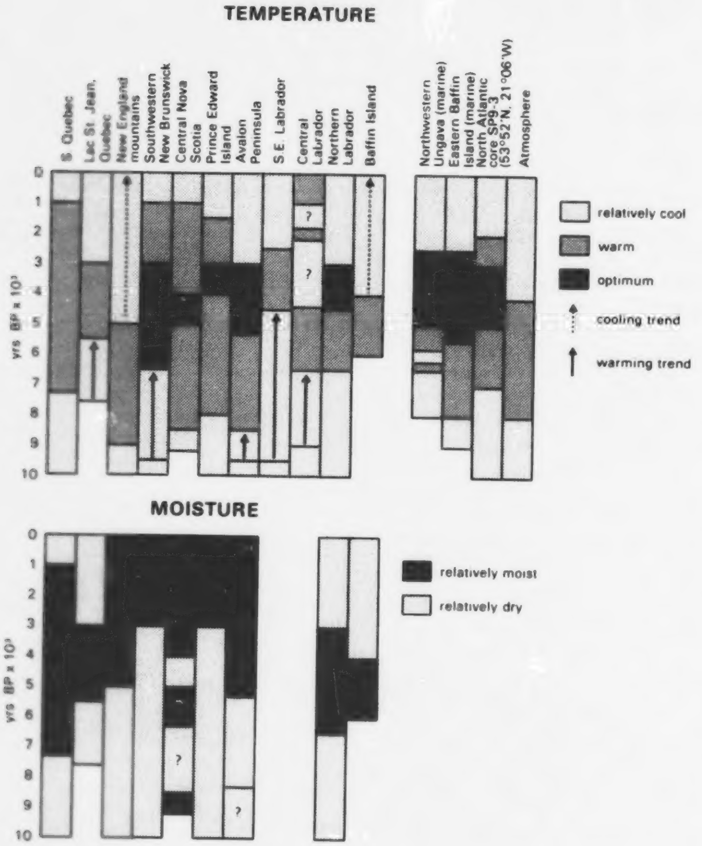


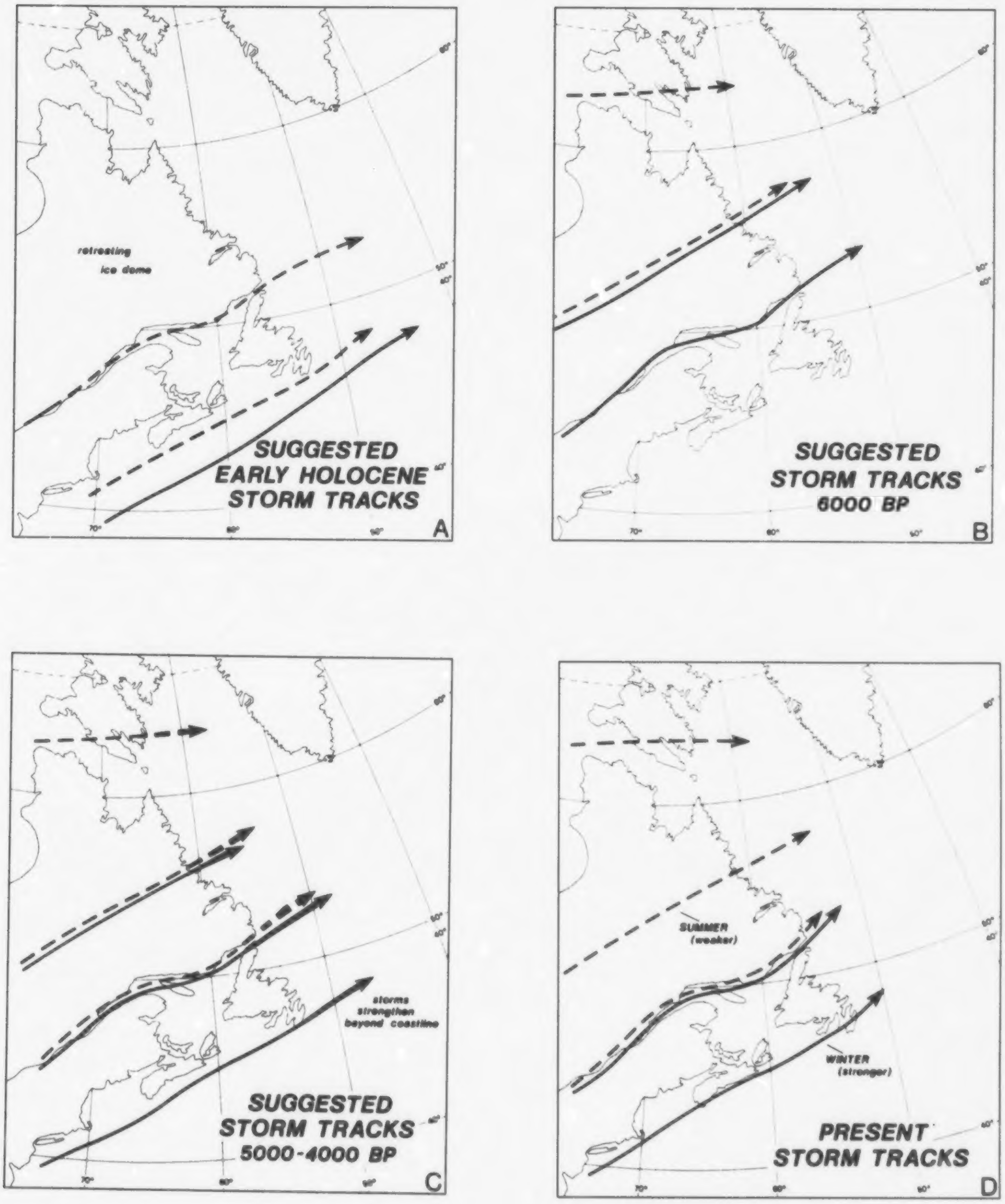
FIGURE 5: *Inferred Holocene climatic conditions, northeastern North America. Sources: S. Québec and Lac St. Jean: Richard (1978, 1980); New England mountains: Davis* et al. *(1980); S.W. New Brunswick: Mott (1975); Central Nova Scotia: Hadden (1975); Pr. Edward Is.: Anderson (1980); Avalon Peninsula: author; S.E. Labrador: Lamb (1980); Central Labrador: Jordan (1975); N. Labrador: Short (1978); Baffin Is.: Andrews* et al. *(1980); N.W. Ungava (marine) and E. Baffin Is. (marine): Andrews (1972); N. Atlantic core SP9-3: Wollin* et al. *(1971); Atmosphere: Dansgaard* et al. *(1971).*

POSSIBLE CONTROLLING FACTORS OF HOLOCENE CLIMATE IN THE REGION

The sequence of climatic changes affecting the Avalon Peninsula can be tentatively ascribed to changes in the balance between the three dominant controls upon the present climate: the winter continental anticyclone over Labrador, the oceanic Bermudan anticyclone and the transient low pressure systems which affect the intervening zone (Banfield 1981). It is these disturbances which, passing across or close to Newfoundland throughout the year, provide abundant precipitation (Figure 6D). In summer, in the absence of the continental anticyclone, the zone affected by low pressure systems extends north to include northern Labrador and Baffin Island. In summer, also, the Bermudan anticyclone, may extend its influence northward to give warm dry conditions in the Maritime Provinces and southern and central Newfoundland.

The margin of the Laurentide ice sheet had begun to withdraw from coastal Labrador by 10,000 yr B.P. (Short and Nichols 1977; Short 1978), and its final disintegration occurred rapidly after 8000 yr B.P. (Bryson *et al.* 1969). The flow of meltwater through Hudson Strait augmented the Labrador Current and delayed the amelioration of water temperatures (Andrews 1972), but by 7000 yr B.P. the oceanic polar front had retreated to its present position south of Greenland from its position south of Newfoundland at 9300 yr B.P. (Ruddiman and Glover 1975). While the ice sheet was extensive, the continental anticyclone would have been more persistent and may even have remained in place throughout the year. Frontal systems would have been steered eastward south of their present tracks (Figure 6A), limiting the advection of heat and moisture to Labrador and Baffin Island, and reducing winter precipitation in Newfoundland, the Maritime Provinces and New England. With the retreat of both the continental ice margin and oceanic polar front, the zone influenced by frontal systems extended northward (Figure 6B), explaining the much later onset of warmer conditions at terrestrial sites in northern Labrador, at about 6500 yr B.P. (Short and Nichols 1977; Short 1978), than in the south of the peninsula where a warming trend is recognized as early as 9000 yr B.P. (Lamb 1980).

The northward shift of the zone affected by frontal systems could well have been accompanied by a greater summer influence of the Bermudan High, maintaining relatively dry conditions over southeastern Canada throughout the period of maximum warmth. By 5000 yr B.P., however, a mechanism other than a simple northward shift of major pressure zones appears to have come into play. The atmospheric optimum was drawing to a close, and a



retreating ice dome
SUGGESTED EARLY HOLOCENE STORM TRACKS
A
SUGGESTED STORM TRACKS 6000 BP
B
storms strengthen beyond coastline
SUGGESTED STORM TRACKS 5000-4000 BP
C
SUMMER (weaker)
WINTER (stronger)
PRESENT STORM TRACKS
D
70°
60°
50°
40°

FIGURE 6: *Holocene storm tracks - Atlantic Canada.*

cooling trend is recognized in the mountains of New England (Davis *et al.* 1980), although evidence for cooling in the presumably less-sensitive lowland regions of interior Québec is lacking (Richard 1978, 1980). At the same time the period of greatest oceanic warmth was just beginning (Wollin *et al.* 1971), as was the warmest period on the Avalon Peninsula; the final warming in Labrador did not occur for another 500 years. Both on the Avalon Peninsula and in Labrador the warmest period was one of increased precipitation. This, combined with evidence of cooling in the more continental area of New England, suggests vigorous cyclone activity for at least part of the year associated with the thermal gradient between the cooling continent and the still-warm ocean (Figure 6C). It was not until 3000 yr B.P., the end of the oceanic temperature optimum, that a cooling trend was registered in the terrestrial record of eastern Canada. The climatic deterioration appears to have been more abrupt in Newfoundland and Labrador than in the Maritime Provinces. Detailed work indicates fluctuations of temperature during the cooler period of the last 3000 years, but the evidence is still too scattered to provide a picture of any general trends that may exist. The cooling at 3000-2500 yr B.P. was accompanied in northern Labrador and Baffin Island by a decrease in precipitation, indicating the increased dominance of dry arctic air masses, and the initiation of the pattern of storm tracks shown in Figure 6D.

SUMMARY

Modern paleobotanical findings by a number of workers in Newfoundland and eastern Labrador-Ungava are synthesized. Present vegetation zones have characteristic pollen spectra, which, although not wholly analogous to fossil spectra, aid in the interpretation of past vegetation. Fossil spectra have been mapped at selected time-horizons from 10,000 yr B.P. permitting the recognition of stages in vegetational development, and the tracing of the immigration of certain taxa. Changing pollen spectra and rates of pollen influx and lacustrine sedimentation provide widespread evidence of optimal conditions from about 6000 yr B.P. to 3000 yr B.P., followed by climatic deterioration. A sequence of changing controls on the regional atmospheric circulation is suggested.

ACKNOWLEDGEMENTS

Figures 1, 2, and 5 appeared in Géographie physique et Quaternaire 36, 1982, and are reproduced by permission of Les Presses de l'Université de Montréal.

My research has been funded by the Natural Sciences and Engineering Research Council of Canada. The figures were drawn in the Memorial University of Newfoundland Cartographic Laboratory.

REFERENCES

Anderson, T.W. 1980. Holocene vegetation and climatic history of Prince Edward Island, Canada. Canadian Journal of Earth Sciences 17:1115-1165.

Andrews, J.T. 1972. Recent and fossil growth rates of marine bivalves, Canadian Arctic, and Late-Quaternary Arctic marine environment. Palaeogeography, Palaeoclimatology, Palaeocology 11:157-176.

Andrews, J.T., W.N. Mode, and P.T. Davis. 1980. Holocene climate based on pollen transfer functions, eastern Canadian Arctic. Arctic and Alpine Research 12:41-64.

Banfield, C.E. 1981. The climatic environment of Newfoundland. In: The Natural Environment of Newfoundland, Past and Present. Edited by: A.G. Macpherson and J.B. Macpherson. Department of Geography, Memorial University. pp. 83-153.

Bartley, D.D. 1967. Pollen analysis of surface samples of vegetation from Arctic Quebec. Pollen et Spores 9:101-105.

Bryson, R.A., W.M. Wendland, J.D. Ives, and J.T. Andrews. 1969. Radiocarbon isochrones on the disintegration of the Laurentide ice sheet. Arctic and Alpine Research 1:1-14.

Dansgaard, W., S.J. Johnsen, H.B. Clausen, and C.C. Langway, Jr. 1971. Climatic record revealed by the Camp Century ice core. In: The Late Cenozoic Glacial Ages. Edited by: K.K. Turekian. Yale University Press, New Haven and London. pp. 37-56.

Davis, M.B., R.W. Spear, and L.C.K. Shane. 1980. Holocene climate of New England. Quaternary Research 14:240-250.

Hadden, K.A. 1975. A pollen diagram from a postglacial peat bog in Hants County, Nova Scotia. Canadian Journal of Botany 53:39-47.

Ives, J.W. 1977. Pollen separation of three North American birches. Arctic and Alpine Research 9:73-80.

Jordan, R. 1975. Pollen diagrams from Hamilton Inlet, central Labrador, and their environmental implications for the northern Maritime Archaic. Arctic Anthropology 12:92-116.

Lamb, H.F. 1980. Late Quaternary vegetational history of southeastern Labrador. Arctic and Alpine Research 12:117-135.

Macpherson, J.B. 1981. The development of the vegetation of Newfoundland and climatic change during the Holocene. In: The Natural Environment of Newfoundland, Past and Present. Edited by: A.G. Macpherson and J.B. Macpherson. Department of Geography, Memorial University. pp. 189-217.

Macpherson, J.B. 1982. Postglacial vegetational history of the eastern Avalon Peninsula, Newfoundland, and Holocene climatic change along the eastern Canadian seaboard. Géographie physique et Quaternaire 36:175-196.

McAndrews, J., et G. Samson. 1977. Analyse pollinique et implications archéologiques et géomorphologiques, Lac de la Hutte Sauvage (Mushuau Nipi), Nouveau-Québec. Géographie physique et Quaternaire 31:177-183.

Mott, R.J. 1974. Modern pollen spectra from Labrador. Geological Survey of Canada Paper 74-1B:232-234.

______. 1975. Palynological studies of lake sediment profiles from southwestern New Brunswick. Canadian Journal of Earth Sciences 12:273-288.

______. 1976. A Holocene pollen profile from the Sept-Iles area, Quebec. Naturaliste Canadien 103:457-467.

Page, G. 1972. The occurrence and growth of trembling aspen in Newfoundland. Department of the Environment, Canadian Forestry Service Publication 1314:1-15.

Railton, J.B. 1973. Vegetational and climatic history of southwestern Nova Scotia in relation to a South Mountain ice cap. Ph.D. thesis, Dalhousie University, Halifax.

Richard, P. 1978. Histoire tardiglaciaire et postglaciaire de la végétation au mont Shefford, Québec. Géographie physique et Quaternaire 32:81-93.

______. 1980. Paléophytogéographie post-wisconsinienne du Québec-Labrador; bilan et perspectives. Département de Géographie, Université de Montréal, Notes et Documents 80-01:1-30.

Rowe, J.S. 1972. Forest regions of Canada. Department of the Environment Canadian Forestry Service Publication 1300. 172 pp.

Ruddiman, W.F., and L.K. Glover. 1975. Subpolar North Atlantic circulation at 9300 yr. BP; faunal evidence. Quaternary Research 5:361-389.

Short, S.K. 1978. Palynology: a Holocene environmental perspective for archaeology in Labrador-Ungava. Arctic Anthropology 15:9-35.

Short, S.K., and H. Nichols. 1977. Holocene pollen diagrams from subarctic Labrador-Ungava: vegetational history and climatic change. Arctic and Alpine Research 9:265-290.

Terasmae, J. 1976. In search of a palynological tundra. Geoscience and Man 15:77-82.

Wollin, G., D.B. Ericson, and M. Ewing. 1971. Late Pleistocene climates recorded in Atlantic and Pacific deep-sea sediments. In: The Late Cenozoic Glacial Ages. Edited by: K.K. Turekian. Yale University Press, New Haven and London. pp. 199-214.

Young, S.B. 1971. The vascular flora of St. Lawrence Island with special reference to the floristic zonation of arctic regions. Harvard University, Contributions of the Gray Herbarium 201:1-115.

LATE-GLACIAL CLIMATIC CHANGE IN THE MARITIME PROVINCES

R.J. Mott[1]

INTRODUCTION

The late-glacial[2], that time during and immediately following the waning of the ice sheets, was a critical period in the Quaternary climatic history of the Maritime Provinces. As land became available, vegetation migrated into the area. This migration was dependent on a number of factors, with climate being one of the critical parameters. Insight into the climatic history can be gained from the vegetational history outlined by palynological study of suitable organic deposits.

Palynological studies in the Maritime Provinces are not as numerous as in some areas of Canada such as Ontario and southern Québec, and relatively few of the sites that have been studied extend far enough back in time to include the late-glacial. However, enough work has been done to outline, in a preliminary way, the vegetational changes, and hence the possible climatic changes, that have affected the area in late-glacial time.

Two types of deposits are important in reconstructing late-glacial paleoenvironments in the Maritimes: buried organic deposits with radiocarbon dates in the 11,000 to 12,000 + yr B.P. range overlain by mineral sediments, and lake-sediment sequences where organic deposition began more than 11,000 years ago and continued to the present. Several recently discovered sites in Nova Scotia, along with some previously reported sites, comprise the list of pertinent buried organic deposits. Six lake-sediment sequences from Nova Scotia and New Brunswick provide the complete pollen profiles of the late-glacial and Holocene. Interestingly, no organic deposits of any kind with radiocarbon dates greater than 10,000 yr B.P. have been found on Prince Edward Island.

---

[1] Geological Survey of Canada, 601 Booth Street, Ottawa, Ontario K1A 0E8

[2] Watts (1980), in his review of late-glacial vegetation and climates of Europe, used 'late-glacial' "... as an informal term to describe the vegetation and inferred climate of the Late Pleistocene interval between approximately 14,000 and 10,000 years ago." The term "late-glacial" is used here in much the same sense.

SITE LOCATION AND DESCRIPTION

Buried Organic Sites

Several sites have been discovered over the years where organic sediments that have produced late-glacial radiocarbon ages are overlain by diamicton or other mineral sediment. In some cases, the diamicton was identified as till or till-like, but in other cases an unequivocal interpretation was not forthcoming. Other sites showed sand of outwash or eolian origin overlying the organic sediment. However, in each case organic sediments had begun to accumulate under apparently suitable climatic conditions, but ceased more or less abruptly with deposition of mineral sediment, suggesting less favourable conditions.

Hickox (1962) was the first to describe such a site on Port Hood Island off the west coast of Cape Breton Island, Nova Scotia (Figure 1). He reported peat dating 10,710 ± 240 yr B.P. (Y-762) between two tills, and postulated that the upper till may have been deposited by ice advancing from a local centre on Cape Breton Highlands. The site was re-examined by Terasmae (1974), and a second date of 11,000 ± 170 yr B.P. (GSC-540) was obtained on woody detritus. A third date of 11,300 ± 160 yr B.P. (GSC-541) on a lower sample of willow (*Salix* sp.) twigs was obtained later (Lowdon and Blake 1976). The dated horizons are organic seams within a 75-cm thick layer of sand overlying reddish, stony, silty clay till, and overlain by buff, brown to reddish, stony, silty clay diamicton. Terasmae concluded that the overlying diamicton was not till, but attributed it to soil creep from unstable slopes induced by a deteriorating climate sometime between 11,000 and 10,000 years ago.

MacNeill (1969) reported a peat layer with an age of 11,670 ± 170 yr B.P. (I-3234) overlain by a till from a site at Benacadie Point, East Bay, Bras D'Or Lake, Cape Breton Island. Grant also sampled an organic layer below till-like diamicton at Benacadie Point (Figure 1) that gave an age of 11,300 ± 90 yr B.P. (GSC-2146; Lowdon and Blake 1976). He found 30cm of twiggy moss peat overlying sand and covered by 2m of till-like diamicton, and he suggested that if the diamicton was not till it might be solifluction debris related to a climatic detrioration equivalent to the Younger Dryas of Europe (Grant in Lowdon and Blake 1976).

Grant also discovered a peat layer at Campbell along the southwest coast of Cape Breton Island (Figure 1). There, approximately 30cm of sandy peat and sand with peat stringers overlies red clay till and is overlain by 60cm of sand of possible outwash origin. A radiocarbon date on the peat produced an age of 11,200 ± 110 yr B.P. (GSC-2212; personal communication, D. Grant, 1982).

Black muck in the upper part of a granitic sand unit about 25-cm thick was discovered by Nielsen (1976) near Nictaux Falls in the Annapolis Valley (Figure 1). The alluvium containing the organic matter overlies red till and is overlain by 30cm of red, clayey, till-like material and 2.7m of red silty sand. The basal 2cm of muck produced a radiocarbon date of 11,200 ± 100 yr B.P. (GSC-2062; Lowdon and Blake 1975).

During the course of mapping soils in the Brookside area northwest of Truro (Figure 1), organic detritus and two organic layers within a 1-m thick sand unit were uncovered (Beke *et al.*, in preparation). The organic-bearing unit was overlain by 1.7m of reddish brown sandy and silty diamicton, and was probably underlain by till similar to that outcropping in the surrounding area. An age of 11,100 ± 100 yr B.P. (GSC-2930) was obtained on the uppermost organic layer (Beke *et al.*, in preparation).

Recent work by Stea has yielded two buried organic sites. The L.E. Shaw Brickyard at Lantz (Figure 1) has an organic layer with a radiocarbon date of 11,100 ± 100 yr B.P. (GSC 3116; Blake 1982). The organic material is in the form of pods and stringers up to 10-cm thick in a buff-gray sand unit. About 1.3m of grey and red clay diamicton overlies the sand. Brown sand 150-cm thick over reddish-brown laminated clay underlies the organic unit. The overlying diamicton is regarded as glacial till or soliflucted mud related to a climatic deterioration that occured after 11,000 yr B.P. (Stea and Grant, in Blake 1982).

The second site investigated by Stea is on the Chignecto Bay coast at Joggins (Figure 1). A 10-15cm thick compact fibrous peat overlies up to 50cm of reddish-brown diamicton and more than 2m of yellow sand with red silt lenses. Overlying the peat is 30-40cm of fine sand of undetermined origin. The radiocarbon date on the peat is 11,400 ± 100 yr B.P. (GSC-3550; personal communication, R. Stea, 1983).

The Big Brook site (Figure 1) in the Georgia-Pacific Gypsum Quarry, about 7km south of River Denys, Cape Breton Island, produced a date of 11,000 ± 90 yr B.P. (GSC-3378), in addition to much older dates from other buried nonglacial intervals. The date was derived from a layer of black, silty organic clay less than 10-cm thick overlying about 2cm of grey clay with minor organics over purplish-red till. About 40cm of grey clay that grades into 50cm of brown clay overlies the organic layer. The modern soil profile is developed in the upper brown clay.

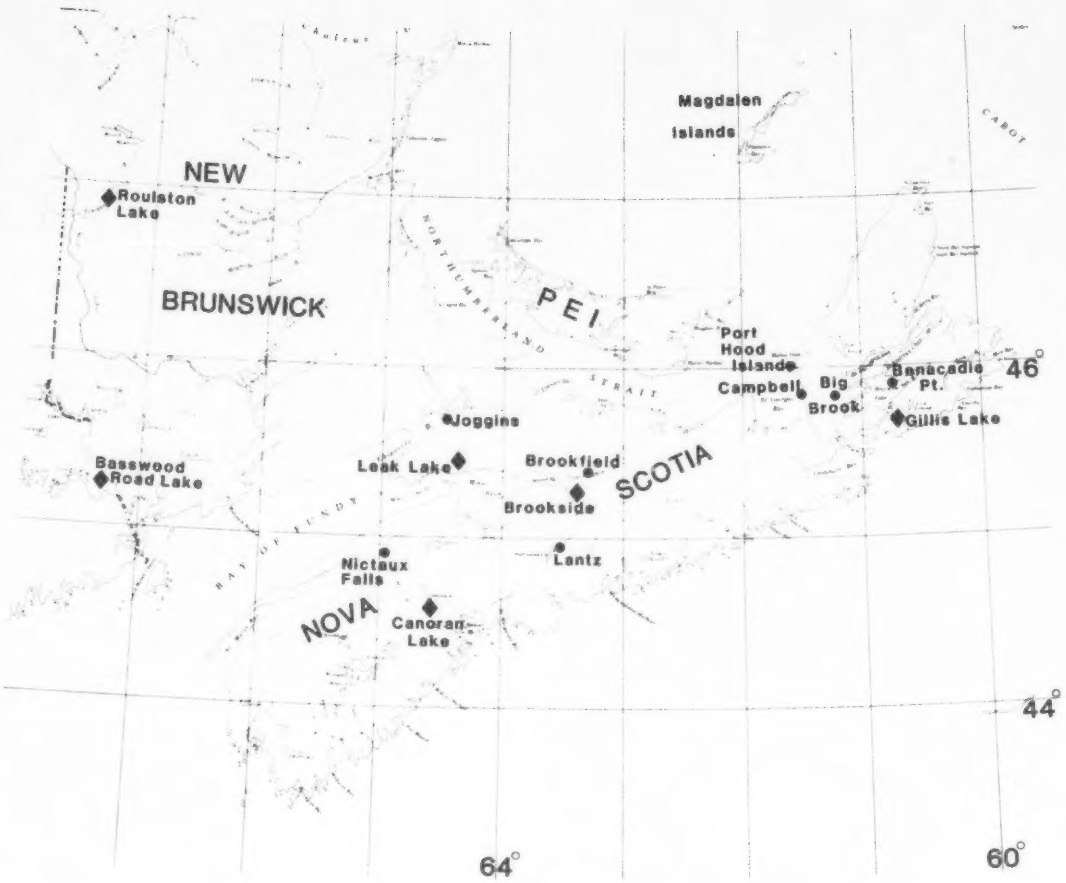

Magdalen
Islands
CABOT
NEW
BRUNSWICK
Roulston
Lake
NORTHUMBERLAND
STRAIT
PEI
Port
Hood
Islands
Campbell
Big
Brook
Benacadie
Pt.
Gillis Lake
46°
Joggins
Basswood
Road Lake
Leak Lake
Brookfield
Brookside
SCOTIA
BAY OF FUNDY
Nictaux
Falls
Lantz
NOVA
Canoran
Lake
44°
64°
60°

FIGURE 1: *Site location map.*

In his study of Quaternary deposits along East Bay, Bras D'Or Lake, Cape Breton Island, Occhietti has discovered several sites with organic sediments overlain by diamicton or other mineral sediment. Preliminary radiocarbon dating has produced ages in the late-glacial range (personal communication, S. Occhietti, 1982). Palynological work has not yet begun on these sites.

Lake-Sediment Sites

Early palynological research in Nova Scotia was done by Livingstone and Livingstone, who in 1958 reported results of work on a core from Gillis Lake, Cape Breton Island (Figure 1), one of the relatively few lake sites found thus far that includes the late-glacial. They found a basal pink diamicton overlain by 2m of grey and greenish-grey laminated clay, 10cm of pinkish-grey clay and 17cm of grey laminated clay similar to the lower grey clay. Brown and grey-brown gyttja covered the clay sequence. The basal gyttja was dated at 10,340 ± 200 yr B.P. (Y-524; Livingstone and Livingstone 1958). Unfortunately, a reliable radiocarbon date on the underlying clay sediments could not be obtained. Several other sites studied by Livingstone and colleagues do not show this sedimentary sequence, and have younger dates at the base (Livingstone 1968; Livingstone and Estes 1967).

Only one of three lake sites from central Nova Scotia studied by Railton (1973) produced a date older than 10,000 yr B.P. The basal organic sediments at Canoran Lake (Figure 1) dated 11,700 ± 160 yr B.P. (GSC-1486). The basal sand in the core was overlain by clay, and in turn by a sequence of algal gyttja broken only by a layer of clay gyttja about 15-cm thick 15cm above the base of the algal gyttja.

Mott (1975) reported on Basswood Road Lake site in southwestern New Brunswick (Figure 1) where an 80-cm thick layer of grey clay occurs within the algal gyttja about 28cm above its base. Grey laminated clay underlies the organic sediments. A date on the base of the gyttja was 12,600 ± 270 yr B.P. (GSC-1067), whereas the upper grey clay was bracketed by dates of 11,300 ± 180 yr B.P. (GSC-1645) and 9460 ± 200 yr B.P. (GSC-1643).

Leak Lake, north of Minas Basin in Nova Scotia (Figure 1), was cored by Wightman for radiocarbon dating and palynological study (Wightman 1980). The basal metre of core showed red silt overlain by about 35cm of grey, slightly organic silt and 40cm of grey clay, overlain in turn by clayey, silty gyttja. A radiocarbon date at the base of the upper clayey gyttja was 9715 ± 200 yr B.P. (DAL-314), and appears reliable judging by the palynology. Anamolous dates of 12,900 ± 160 (GSC-2728) and 15,900 ± 1200 yr B.P. (GSC-2880) were obtained from adjacent increments in the lower part of the basal organic silt (Wightman 1980). The dates may be spurious as a result of contamination by old carbon derived from Carboniferous bedrock of the area.

More recently, the author has collected cores at other sites in New Brunswick. One site in particular, Roulston Lake near Plastic Rock (Figure 1), is of interest because of a band of silty clay about 30cm above the base of the lower silty-clayey gyttja. Above the silty clay layer is silty gyttja and algal gyttja. A date of 11,100 ± 90 yr B.P. (GSC-2804) was obtained from the basal organic sediment, and a date of 9930 ± 160 yr B.P. (GSC-2872) was obtained from the organic sediment immediately above the silty clay layer.

A section through peat, organic lake sediments and laminated clay, exposed during overburden removal at the Maritime Cement, Canada Cement Lafarge Quarry near Brookfield, Nova Scotia (Figure 1), is interesting in that it occurs about mid-way between the sites at Brookside and Lantz, both of which have diamicton overlying organic horizons. However, at Brookfield sedimentation appears continuous through basal laminated (varved?) clay into clayey marl lake sediment 50-cm thick, silty clay 50-cm thick, 45cm of algal gyttja and 60 cm or more of peat. Unfortunately, the basal marl precludes dating because of the probability of contamination by old carbonates. A wood fragment from about 10cm above the base of the algal gyttja gave an age of 9140 ± 170 yr B.P. (GSC-3635). Algal gyttja from the same interval produced a date of 9780 ± 90 yr B.P. (GSC-3652) showing a discrepancy of several hundred years between sediment and enclosed wood[1], an indication of contamination by old carbonate at the site. Despite the lack of basal dates to provide a precise chronology, the sequence appears to extend back into the late-glacial, and useful information on the climate can be provided by pollen and micro- and macrofossil analyses.

---

[1] The wood was found lying along the bedding plane of the sediment, and showed no sign of having penetrated into the existing sediment to any significant extent at time of deposition.

PALYNOLOGICAL RESULTS

Buried Organic Sites

Preliminary palynological results from several buried organic sites are included in Figure 2 to illustrate the changes that occurred. The diagram is arranged as a general transect from southwest to northeast with Nictaux Falls Site shown at the base and the Port Hood Isand Site at the top. The latter is an abbreviated form of the diagram reported by Terasmae (1974), and the percentages are based on total arboreal pollen. Total NAP (non arboreal pollen) is shown in the Cyperaceae column. Percentages from all other sites are based on total palynomorphs exclusive of Cyperaceae, aquatics and spores. Taxa combined in columns labelled "Other Shrubs" and "Other herbs" include useful climatic indicators, but for this presentation only the gross features of the pollen spectra are shown. A composite sample representing the complete organic increment was used for radiocarbon dating at all sites except Port Hood Island, Brookside and Nictaux Falls where the increments used are shown on the diagram by arrows.

Three sites, Nictaux Falls, Lantz and Brookside, are characterized by spectra dominated by spruce (*Picea*) and sedge (Cyperaceae) with lower amounts of pine (*Pinus*), birch (*Betula*) and willow (*Salix*) pollen. Grass (Gramineae) pollen and fern spores (Polypodiaceae) are present and are particularly abundant at Brookside and Nictaux Falls respectively.

Joggins, Big Brook and Benacadie Point, by contrast, have less spruce and more pine pollen. Sedge pollen is abundant especially at Joggins and Big Brook. Birch and willow are abundant, as are grass pollen and lycopod (Lycopodiaceae) and fern spores at Big Brook and Benacadie Point. Buffalo berry (*Shepherdia canadensis*) has high percentages at Joggins as do *Selaginella selaginoides* spores at Big Brook.

The spectra at Campbell and Port Hood Island are both dominated by birch and willow pollen, with willow decreasing as birch increases. Non arboreal pollen values are high at both sites with sedge pollen especially abundant at the Campbell Site.

Lake Sediment Sites

Spectra from six lake sites are presented in Figures 3 through 8. The Gillis Lake diagram is an abbreviated form of the diagram reported by Livingstone and Livingstone

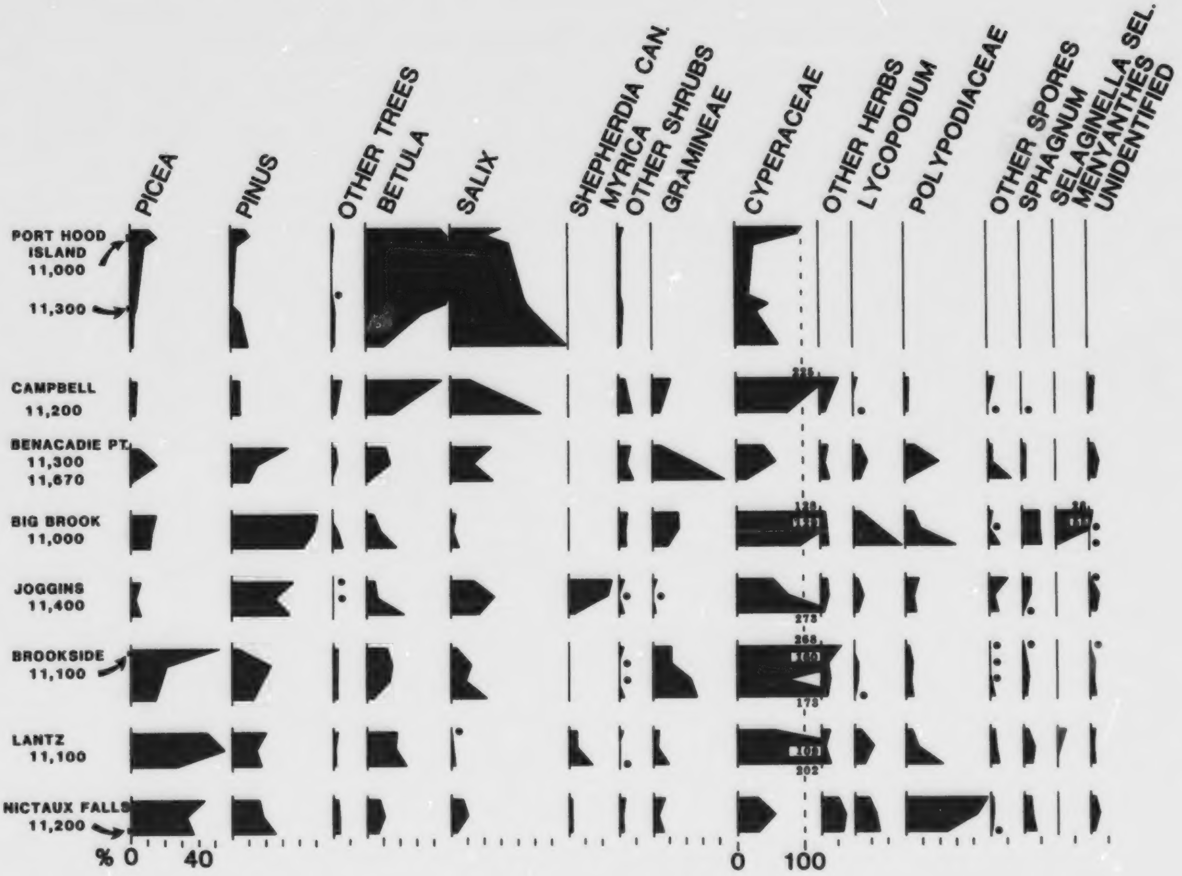


FIGURE 2: *Composite pollen diagram for buried organic sites. Abbreviated Port Hood Island pollen profile is from Terasmae (1974). Abbreviated Brookside pollen profile is from Beke* et al. *(in preparation).*

(1958). Percentages are based on total palynomorphs. Railton's (1973) Canoran Lake diagram has percentages based on total terrestrial pollen, and it also shows only the most prominent taxa. The Leak Lake and Basswood Road Lake diagrams are from Wightman (1980) and Mott (1975), respectively, and along with the Roulston Lake and Brookfield diagrams have percentages based on total pollen exclusive of sedge and aquatics. The different pollen sums certainly affect the percentages values obtained for individual taxa, but not enough to distort the overall trends apparent in the profiles.

Sedimentation began early at Basswood Road Lake (Figure 3) in southwestern New Brunswick where a basal date of 12,600 yr B.P. was obtained (Mott 1975). Sedge, grass, *Artemisia* and willow pollen characterize the basal sediments. This assemblage is followed by birch, poplar/aspen (*Populus*) and spruce pollen assemblage zones prior to 11,300 yr B.P. About 11,000 yr B.P., sedimentation changed abruptly to clay in which spruce pollen was less abundant, and sedge and willow pollen and fern spores were more abundant. Approximately 10,000 years ago, sedge pollen decreased sharply in abundance and birch, spruce and poplar/aspen and pollen of various shrubs increased again as organic deposition became dominant. Pollen influx values that were increased prior to 11,000 yr B.P. declined before increasing again in early Holocene time (after 10,000 yr B.P.).

Farther north in New Brunswick, at Roulston Lake (Figure 4), birch and willow pollen decline after 11,100 yr B.P. and are replaced by a sedge, *Artemisia*, grass and herb dominated assemblage. Before 9930 yr B.P., these latter taxa decline as birch becomes dominant once again. Increasing pollen concentration values of the lowest zone decline between 11,000 and 10,000 yr B.P. and rise again after 10,000 yr B.P.

The Leak Lake diagram (Figure 5) shows a willow, sedge, *Artemisia* pollen assemblage in the basal silty clay and silty gyttja changing to an assemblage dominanted by birch, and then to one dominated by spruce (Mott, unpublished Geological Survey of Canada Palynological Report No. 79-13 in Wightman 1980). Spruce pollen abundance declines sharply in the overlying grey clay where sedge pollen and lycopod and fern spores, along with pine and poplar/aspen pollen, become dominant. These taxa decline again as birch and willow and then spruce characterize the overlying organic sediment. Pollen concentrations, which were increasing in the silty gyttja, decline sharply in the grey clay and increase again dramatically in the upper sediment. Unfortunately, the two basal dates, 12,900 and 15,900 yr B.P. are anomalous and do not provide a useful chronology.

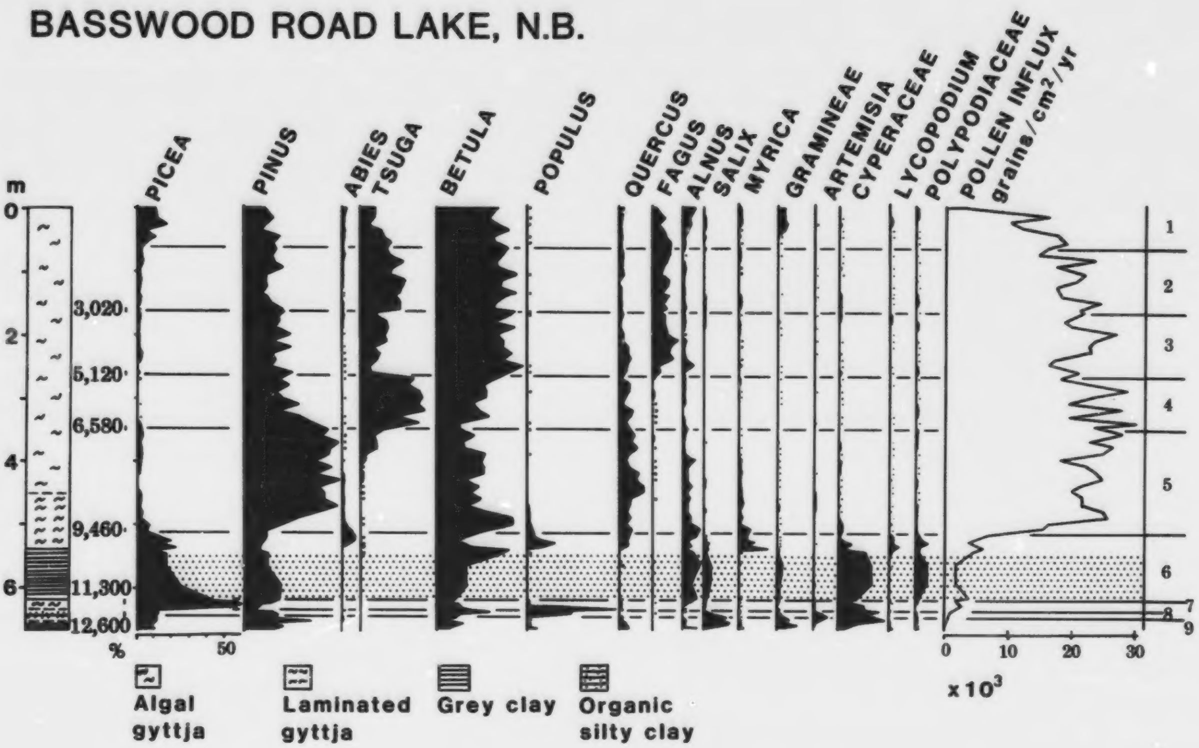


FIGURE 3: *Abbreviated Basswood Road Lake, New Brunswick, pollen diagram from Mott (1975). Shaded pattern in this and following diagrams delineates the pollen spectra of the cool period following initial warming.*

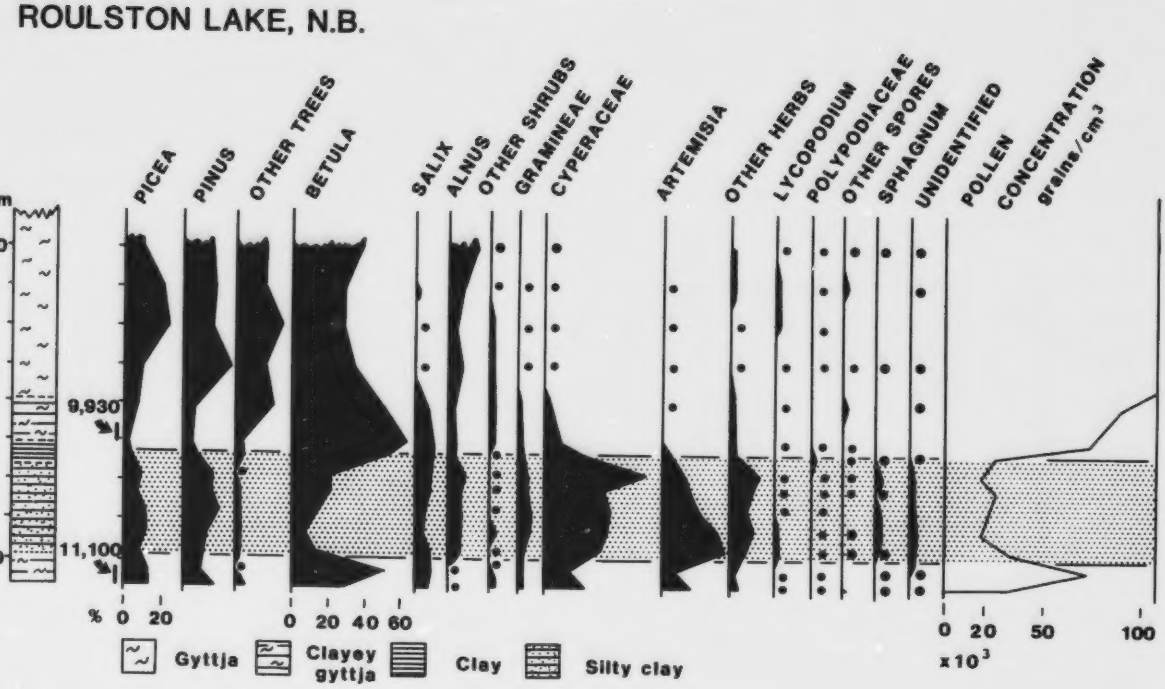


FIGURE 4: *Roulston Lake, New Brunswick, pollen diagram.*

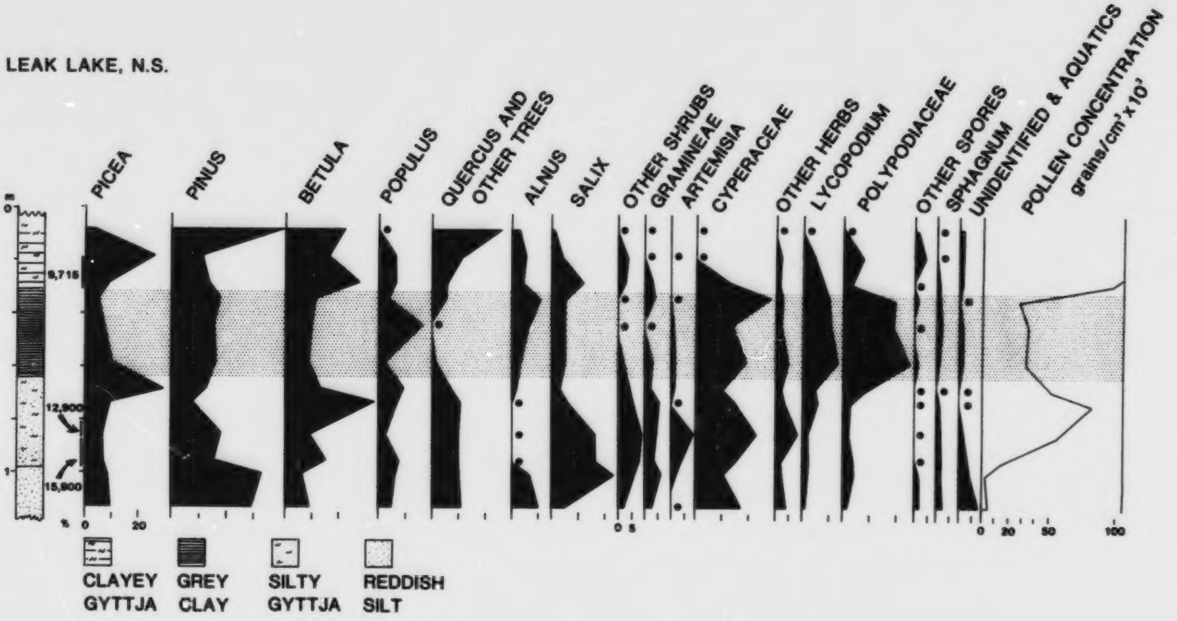


FIGURE 5: *Abbreviated Leak Lake, Nova Scotia, pollen diagram from Wightman (1980).*

The basal sand and clay at Canoran Lake (Figure 6) are essentially devoid of pollen (Railton 1973). At the base of the algal gyttja, birch, *Myrica*, pine and herb pollen are abundant, but give way to a spruce-dominated zone. Spruce declines in the silty gyttja, and birch and sedge pollen and lycopod and fern spores increase. The algal gyttja above the silty gyttja is again dominated by spruce with abundant birch and few shrub and herb pollen types. The date of 11,700 yr B.P. was obtained for the boundary between the birch-*Myrica*-pine zone and the spruce zone making the assemblage contained in the silty gyttja somewhat younger, possibly in the 10,000 to 11,000 yr B.P. range, judging by the overlying abrupt increase in pollen concentrations generally dated at about 10,000 years B.P. at other sites in the Maritimes.

The Brookfield Site (Figure 7) has a birch-sedge pollen assemblage in the basal marly silt and clay, changing to a sedge-birch-*Isoetes* assemblage in the overlying silty clay, and back to a birch-dominated assemblage in the upper organic sediments. Both pollen concentrations and organic content are lower in the silty clay layer. Although radiocarbon control is lacking because of problems associated with dating marly sediments, the date of 9140 yr B.P. on a piece of balsam fir (*Abies balsamea*) wood from above the base of the overlying algal gyttja suggests that the silty clay unit probably dates to more than 10,000 yr B.P.

The Gillis Lake diagram (Figure 8) begins with a pollen zone (L1) dominated by sedge, willow, Ericaceae and pine pollen (Livingstone and Livingstone 1958). Birch pollen predominates in the following zone (L2) along with Ericaceae pollen, and sedge, willow and pine pollen decline. Zone L3 shows a reversion back to less birch and more sedge, willow and pine pollen. About 10,340 years ago birch pollen once again dominates along with spruce pollen in zone A1. Percent loss on ignition indicates the organic content rises through zones L1 and L2, declines in zone L3, and increases greatly in zone A1.

INTERPRETATION

The six lake-site diagrams all show similar trends in vegetation changes, even though they do not all have the same pollen assemblage zones. Pollen content of the early sediments indicates sparse herbaceous vegetation was the first type of vegetation present. Gradually, vegetation cover increased and, in the southwest, trees invaded the region. The vegetation then reverted to more herbs and shrubs and less trees, or more herbs and less

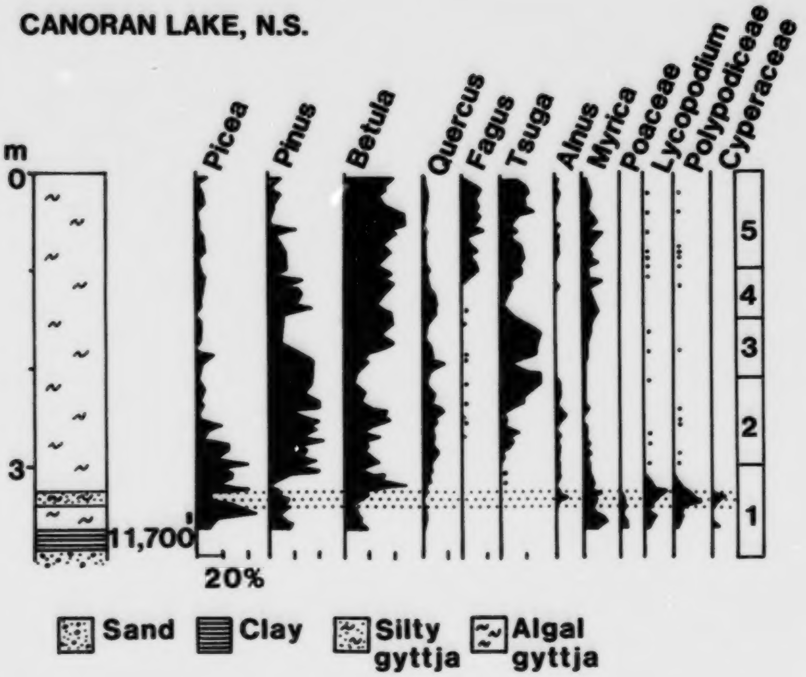


FIGURE 6: *Abbreviated Canoran Lake, Nova Scotia, pollen diagram from Railton (1973).*

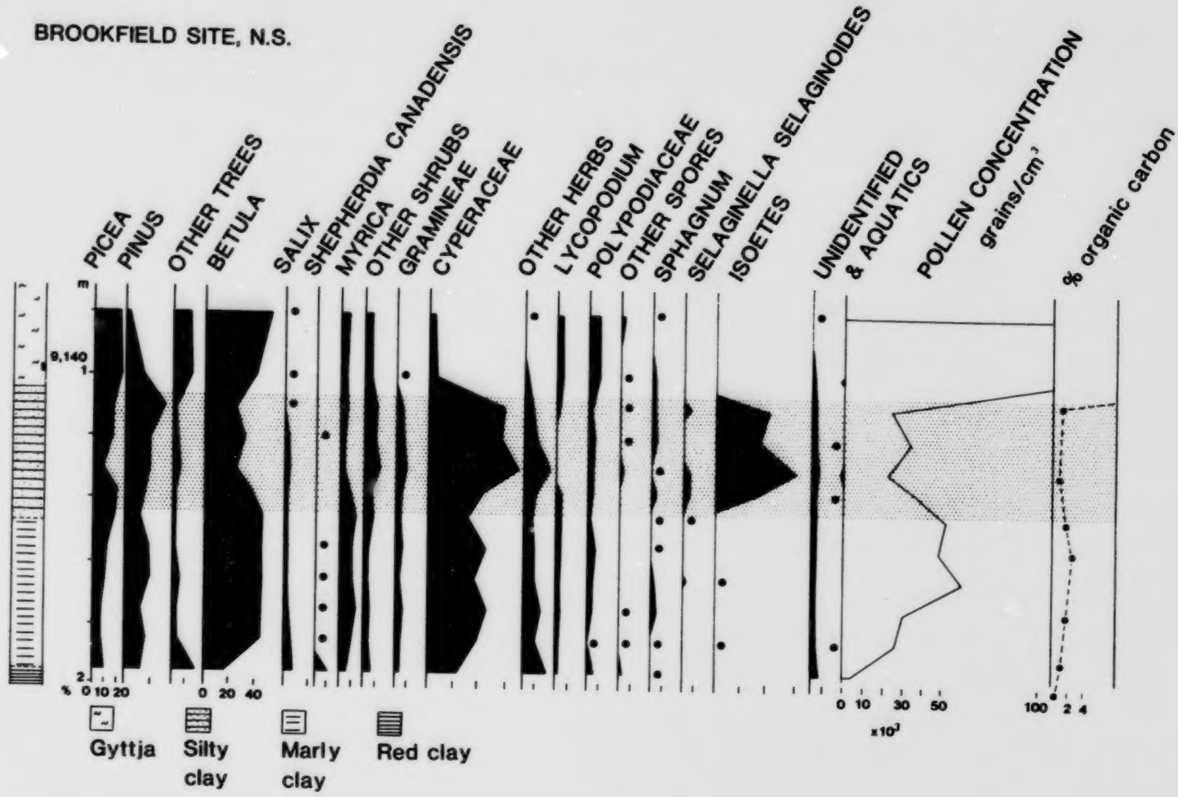


FIGURE 7: *Brookfield site, Nova Scotia, pollen diagram.*

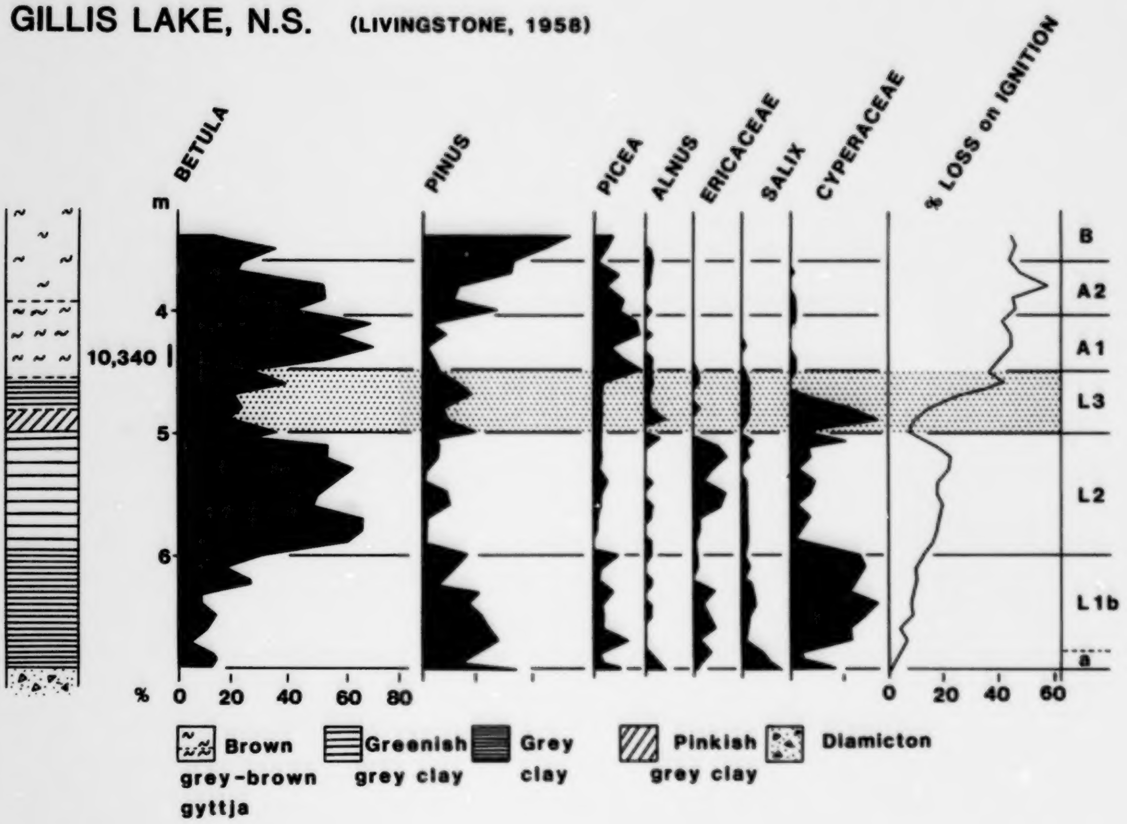


FIGURE 8: *Abbreviated Gillis Lake, Nova Scotia, pollen diagram from Livingstone and Livingstone (1958).*

shrubs depending on the previous vegetation type. In southwestern New Brunswick, what was probably a spruce woodland changed to a vegetation cover with less spruce and more shrubs and herbs. Similarly, in southwestern Nova Scotia, particularly at Canoran Lake and Leak Lake, spruce declined in abundance and more shrubs and herbs appeared. Spruce trees had not reached the Roulston Lake area of New Brunswick, but the birch shrub tundra type of environment reverted to herbaceous tundra. The birch shrub-dominated tundra of Cape Breton Island deteriorated to herbaceous tundra similar to what had existed much earlier. All lake sites then show a proliferation of vegetation with more trees and shrubs and less herbs.

The buried organic sites tell a similar story. Open spruce woodlands with abundant shrubs and herbs existed in southwestern Nova Scotia. Spruce appears to have been migrating to the northeast as time progressed. Shrub tundra or shrub/herb tundra existed in Cape Breton Island. Although pine pollen is abundant at some sites, pine trees were probably not present. The high pine pollen values represent long-distance transport and overrepresentation in the low local pollen abundance of tundra areas. Then, at all sites organic sedimentation ceased, and mineral sediments were deposited.

Radiocarbon dates outline the chronology of events. Vegetation had migrated into southwestern New Brunswick prior to 12,000 yr B.P. In fact, all vegetation changes are earliest at the Basswood Road Lake site suggesting the New England/New Brunswick coastal area was the major early migration route for plants into the Maritimes. By at least 11,700 yr B.P., and possibly much earlier, tundra vegetation was widespread in Nova Scotia and New Brunswick, and by 11,000 yr B.P. or slightly earlier spruce, and probably other trees, had invaded southwestern New Brunswick and much of mainland Nova Scotia. At 11,000 yr B.P., or slightly later, the vegetation changed. Trees became less abundant and shrub and herb populations proliferated. Herb-dominated communities replaced shrub communities in the northeast, and at some sites the vegetation was decimated. These changed conditions persisted until about 10,000 yr B.P., or slightly earlier, when vegetation began to revert to previous types, and after 10,000 yr B.P. vegetation proliferated in all areas.

Whether or not the cessation of organic accumulation and the onset of mineral sedimentation can be related to renewed ice build-up, or reactivation of any existing ice, is not known at present. Possibly, these sedimentation changes can be accounted for simply by the reactivation of slope processes in response to changing climate. The lake sites were certainly not overrun by ice, and the renewed mineral sedimentation can readily be accounted for by solifluction.

CONCLUSIONS AND DISCUSSION

The combined evidence from the lake sites and buried organic sites allows a preliminary reconstruction of climatic changes during late-glacial time in the Maritimes. Plants migrated into the region as deglaciation progressed, earliest in southwestern New Brunswick where there is a date of 12,600 yr B.P. As the climate warmed, vegetation cover increased and spruce trees migrated to the northeast until, by 11,700 yr B.P. or earlier, spruce woodland existed in southwestern Nova Scotia. Arboreal vegetation was even more widespread in southwestern New Brunswick. Shrub and herb tundra environments were characteristic of northeastern Nova Scotia including Cape Breton Island. The climate deteriorated abruptly after 11,000 yr B.P., and the vegetation responded. Trees became less abundant in areas where trees were present, and shrub and herb tundra areas increased. Solifluction increased curtailing organic accumulation at some sites, but there is no firm evidence as yet for reactivation of any remnant ice or renewed accumulation of ice. This cooler period lasted for several hundred years until about 10,000 yr B.P. or slightly earlier, when the climate began to warm again.

I am tempted to relate the warming trend of 12,000 to 11,000 years ago followed by a cooler interval between 11,000 and 10,000 yr B.P., to the Allerød/Younger Dryas climatic oscillation of Europe, as others have done in the past for the apparent climatic oscillation recorded at individual sites in the Maritimes (Livingstone and Livingstone 1958; Grant, in Lowdon and Blake 1976). These speculations, like others for sites throughout northeastern North America, were not widely accepted because the evidence could be refuted, or was found unconvincing (Mercer 1969; McDonald 1971; Wright 1971). Nevertheless, the preliminary evidence presented here, together with evidence from Newfoundland (Grant 1969; Anderson 1983) and unpublished evidence from the Magdalen Islands, Québec, indicate a significant event occurred during late-glacial time. *An event of such magnitude was probably climatic.*

Study of North Atlantic deep-sea cores have shown that deglacial warming began about 13,500 yr B.P., and this warming was manifested in the gradual retreat of the Arctic Polar Front from its near full glacial east-west position at about 40°N., to a position near Greenland and Iceland about 9200 yr B.P. (Ruddiman and McIntyre 1973, 1977, 1981; Ruddiman *et al.* 1977). One reversal interupted this deglacial warming when the Polar Front moved southeasterly once again to near its full glacial position about 10,200 to 10,400 yr B.P. (Ruddiman and McIntyre 1973). The movement of the Polar Front, which acted as a line hinged in the western North Atlantic southeast of Newfoundland sweeping in an arc across the eastern North Atlantic, can be related to the Allerød warming and Younger Dryas cooling of Europe (Ruddiman and McIntyre 1977). The relatively stable position of the Polar Front in the western North Atlantic has been cited to account for the apparent lack of an equivalent synchronous climatic oscillation in northeastern North America (Ruddiman and McIntyre 1973, 1981). However, the Arctic Polar Front may not have been as stable as suspected in the western North Atlantic and northeastern North America, and may have a significant influence on the climate of the Maritimes.

More paleoecological work is needed to outline the climate changes of this critical late-glacial period, when flora and fauna began migrating into the Maritime Provinces. Further palynological work, as well as analyses of various micro- and macrofossils and other parameters is required to characterize the climate changes. Detailed radiocarbon dating of the upper and lower boundaries of buried organic layers, and of mineral-sediment boundaries in basal lake sediments sequences is required to delineate the chronology before the complete significance of late-glacial climatic events is understood.

REFERENCES

Anderson, T.W. 1983. Preliminary evidence for Late Wisconsinan climatic fluctuations from pollen stratigraphy in Burin Peninsula, Newfoundland. Geological Survey of Canada Paper 83-1B:185-188.

Beke, G.L., D.R. Grant, R.J. Mott, and J.A. McKeague. (In preparation). Stratigraphy and significance of a buried Late Wisconsinan organic bed at Brookside, Nova Scotia.

Blake, W., Jr. 1982. Geological Survey of Canada Radiocarbon Dates XXII. Geological Survey of Canada Paper 82-7:1-22.

Grant, D.R. 1969. Surficial deposits, geomorphic features, and Late Quaternary history of the terminus of the Northern Peninsula of Newfoundland and adjacent Quebec-Labrador. Maritime Sediments 5(3):123-125.

______. 1975. Surficial geology of northern Cape Breton Island. Geological Survey of Canada Paper 75-1A:407-408.

Hickox, C.F., Jr. 1962. Late Pleistocene ice cap centered on Nova Scotia. Geological Society of America Bulletin 73:505-510

Livingstone, D.A. 1968. Some interstadial and postglacial pollen diagrams from eastern Canada. Ecological Monographs 38(2):87-125.

Livingstone, D.A., and A.H. Estes. 1967. A carbon-dated pollen diagram from the Cape Breton Plateau, Nova Scotia. Canadian Journal of Botany 45:339-359.

Livingstone, D.A., and B.G.R. Livingstone. 1958. Late-glacial and postglacial vegetation from Gillis Lake in Richmond county, Cape Breton Island, Nova Scotia. American Journal of Science 256:341-359.

Lowdon, J.A., and W. Blake, Jr. 1975. Geological Survey of Canada Radiocarbon Dates XV. Geological Survey of Canada Paper 75-7:1-32.

Lowdon, J.A., and W. Blake, Jr. 1976. Geological Survey of Canada Radiocarbon Dates XVI. Geological Survey of Canada Paper 76-7:1-21.

MacNeill, R.H. 1969. Some dates relating to dating of the last major ice sheet. Maritime Sediments 5(1):1-3.

McDonald, B.C. 1971. Late Quaternary stratigraphy and deglaciation in Eastern Canada. In: The Late Cenozoic Glacial Ages. Edited by: K.K. Turekian. Yale University Press, New Haven and London. pp. 331-353.

Mercer, J.H. 1969. The Allerød oscillation: a European climatic anomaly? Arctic and Alpine Research 1(4):227-234.

Mott, R.J. 1975. Palynological studies of lake sediment profiles from southwestern New Brunswick. Canadian Journal of Earth Sciences 12(2):273-288.

Nielsen, E. 1976. The composition and origin of Wisconsinan till in mainland Nova Scotia. Ph.D. thesis, Dalhousie University, Halifax. 257 pp.

Railton, J.B. 1973. Vegetational and climatic history of southwestern Nova Scotia in relation to a South Mountain ice cap. Ph.D. thesis, Dalhousie University, Halifax. 146 pp.

Ruddiman, W., and A. McIntyre. 1973. Time-transgressive deglacial retreat of polar waters from the North Atlantic. Quaternary Research 3:117-130.

_______. 1977. Late-Quaternary surface ocean kinematics and climatic change in the high-latitude North Atlantic. Journal of Geophysical Research 82:3877-3887.

_______. 1981. The North Atlantic Ocean during the last glaciation. Palaeogeography, Palaeoclimatology, Palaeoecology 35:145-214.

Ruddiman, W., C.D. Sancetti, and A. McIntyre. 1977. Glacial/interglacial response rate of subpolar North Atlantic water to climatic change: the record in ocean sediments. Philosophical Transactions of the Royal Society (London), B, 280:119-142.

Terasmae, J. 1974. Deglaciation of Port Hood Island, Nova Scotia. Canadian Journal of Earth Sciences 11(10):1357-1365.

Watts, W.A. 1980. Regional variation in the response of vegetation to late-glacial climatic events in Europe. In: Studies in the Late-glacial of Northwest Europe. Edited by: J.J. Lowe, J.M. Gray and J.E. Robinson. Pergamon Press pp. 1-21.

Wightman, D.M. 1980. Late Pleistocene glaciofluvial and glaciomarine sediments on the north side of the Minas Basin, Nova Scotia. Ph.D. thesis, Dalhousie University, Halifax. 426 pp.

Wright, H.E., Jr. 1971. Late Quaternary vegetational history of North America. In: The Late Cenozoic Glacial Ages. Edited by: K.K. Turekian. Yale University Press, New Haven and London. pp. 425-464.

MEAN JULY TEMPERATURE AT 6000 YR B.P. IN EASTERN NORTH AMERICA: REGRESSION EQUATIONS FOR ESTIMATES FROM FOSSIL-POLLEN DATA

Patrick J. Bartlein[1] and Thompson Webb, III[2]

INTRODUCTION

The testing of climate-model simulations for past climates requires subcontinental to global maps of paleoclimatic estimates. Such maps reveal the magnitude and pattern of temperature and precipitation over enough model grid-points that useful comparisons are possible. CLIMAP (1976) pioneered in quantitative paleoclimatic reconstruction at a global scale for 18,000 yr B.P., and Kellogg (1978), Street and Grove (1979), Peterson *et al.* (1979), and Butzer (1980) have also produced global maps with paleoclimatic summaries. Recent modeling of Holocene climates requires syntheses of Holcene data at comparable scales (Webb 1984), and we therefore assembled over 200 radiocarbon-dated pollen diagrams from eastern North America as part of the COHMAP (Cooperative Holocene Mapping Project) effort to map global-scale patterns in Holocene climates. The climatic calibration of these pollen data for 6000 yr B.P. required use of 13 regression equations, each from a different subregion of eastern North America. Within this article, we have described the calculation of these equations, as well as the map of estimated mean July temperatures for 6000 yr B.P., that these equations were used to produce.

METHODS

Palynologists have used several statistical procedures in the climatic calibration of pollen data. The procedures include multiple regression analysis (Howe and Webb 1977, 1983), canonical correlation (Webb and Bryson 1972), and principal components analysis followed by multiple regression (Cole 1969; Kay 1979; Heusser and Streeter 1980; Andrews *et al.* 1980). Webb and Clark (1977) described similarities among these methods and concluded that multiple regression was the simplest and most straightforward procedure to use. It requires relatively few assumptions about the statistical properties of the pollen

---

[1] Department of Geography, University of Oregon, Eugene, Oregon 97403-1846

[2] Department of Geological Sciences, Brown University, Providence, Rhode Island 02912-1846

and climate data, and computer programs exist that allow the analysis to proceed by a series of steps in which the statistical assumptions can be checked (Arigo *et al.* 1982; Bartlein and Webb 1984A). The analysis starts with maps and scatter diagrams of the pollen and climatic data, and ends with the calculation of a multiple regression equation between a single climate variable and several pollen variables. Howe and Webb (1983) and Bartlein and Webb (1984B) described the statistical reasons for each step in the analysis and presented detailed examples of their use.

MODEL AND ASSUMPTIONS

The multiple regression model used for climatic calibration has the form:

$$[1] \qquad \underset{n\sim 1}{Y} = \underset{n\sim m\sim 1}{X\beta} + \underset{n\sim 1}{\varepsilon}$$

where n is the number of samples

m is the number of pollen types,

$\underset{n\sim 1}{Y}$ is a n x 1 vector of values for a particular climate variable,

$\underset{n\sim m}{X}$ is a n x m matrix of pollen percentages,

$\underset{n\sim 1}{\beta}$ is a m x 1 vector of regression coefficients, and

$\underset{n\sim 1}{\varepsilon}$ is an n x 1 vector of errors.

The estimation of the regression coefficients by ordinary least squares (Bartlein and Webb 1984B) requires four main assumptions:

(1) that the errors, $\underset{\sim}{\varepsilon}$, are independent, identically normally distributed random variables with a mean of zero, and a variance of σ^2;

(2) that the pollen variables used as predictors, $\underset{\sim}{X}$, have been observed without error;

(3) that the pollen variables used as predictors are neither perfectly or highly correlated with one another; and

(4) that the model has been correctly specified, i.e. that the right predictors have been included, and that the relationship between the climate variable and the predictors is linear.

Bartlein and Webb (1984B) described why these assumptions arise, what the effects of violating them are on the analysis, how to test for violations, and steps to take to assure that a final equation satisfies the assumptions (Table 1). We have found that model specification errors (violations of assumption 4) are more likely to arise in practice than violations of the other assumptions. These errors arise from several sources including: (a) selection of too large or too small a geographic region containing the modern data used in the regression analysis; (b) neglect of nonlinear relationships between the climate and pollen variables; and (c) omission of potentially useful predictors from the model. When specification errors are minimized, other sources of assumption violations are generally minimized as well.

DATA

Modern Pollen and Climatic Data

The modern pollen data came from over 1700 samples of surficial sediments from mostly lakes and bogs, but also from some reservoirs and estuaries (Figure 1). Delcourt *et al.* (1984) have provided a bibliography for most of the published data and also mapped the broad spatial patterns of the 24 most abundant pollen types. Maps of the herb pollen types appear in Birks *et al.* (1975), Webb and McAndrews (1976), and Bernabo and Webb (1977). The modern climatic data used with each pollen sample came from the closest climatic station. South of 50°N, where the climatic stations are dense, only one pollen sample was used per climatic station, but north of 50°N, where the climatic stations are sparse, we used the pollen data from all locations, and thus had data from several pollen sites associated with the climatic data from single meteorological stations.

Pollen Data for 6000 yr B.P.

There were 211 sites with pollen data dated to 6000 yr B.P. (Figure 2, Table 2[1]). The procedures used to gain the interpolated pollen percentages at 6000 yr B.P. are described in Webb *et al.* (1983) and Webb *et al.* (1984). The steps include:

---

[1] Located at end of paper.

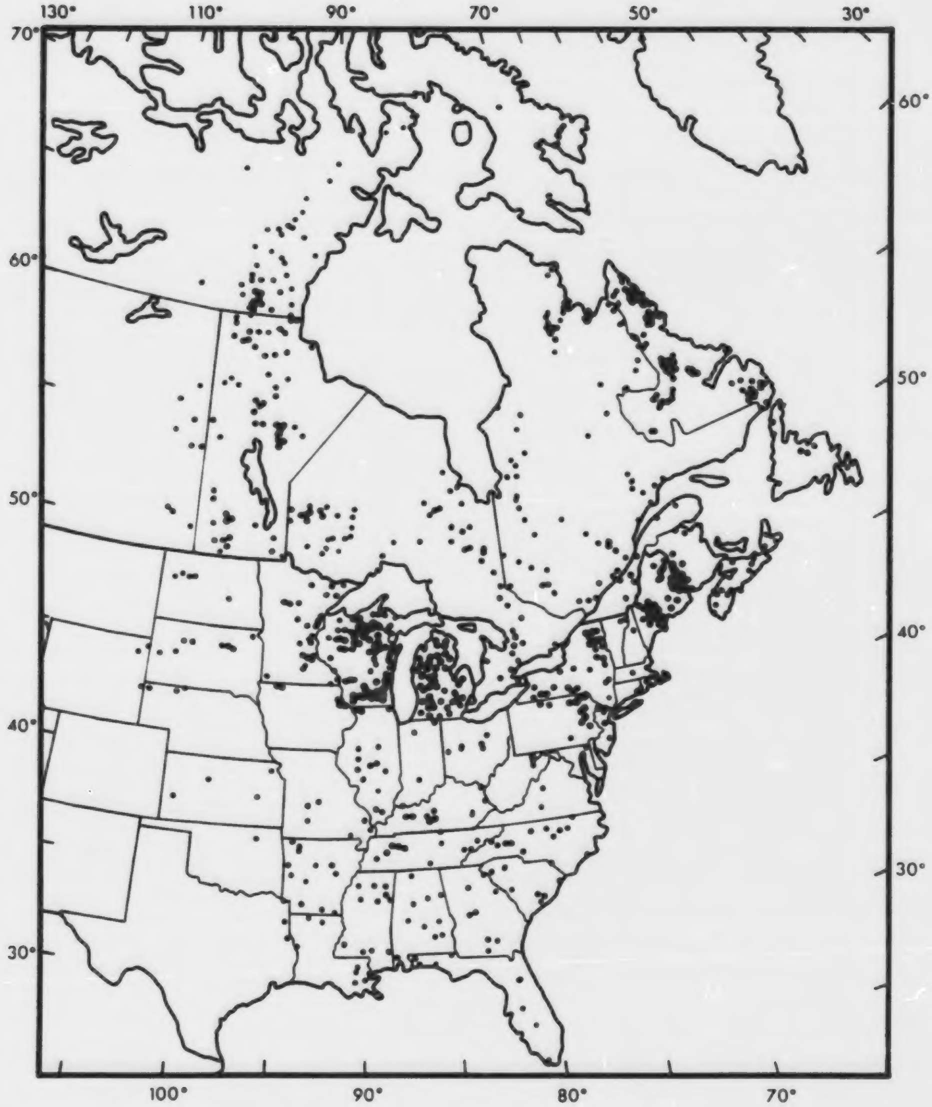

130°
110°
90°
70°
50°
30°
70°
60°
50°
40°
30°
60°
50°
40°
30°
100°
90°
80°
70°

FIGURE 1: *Sites with modern pollen data. Most sites are lakes, but data also came from bogs, cattle-watering tanks, reservoirs, and moss polsters.*

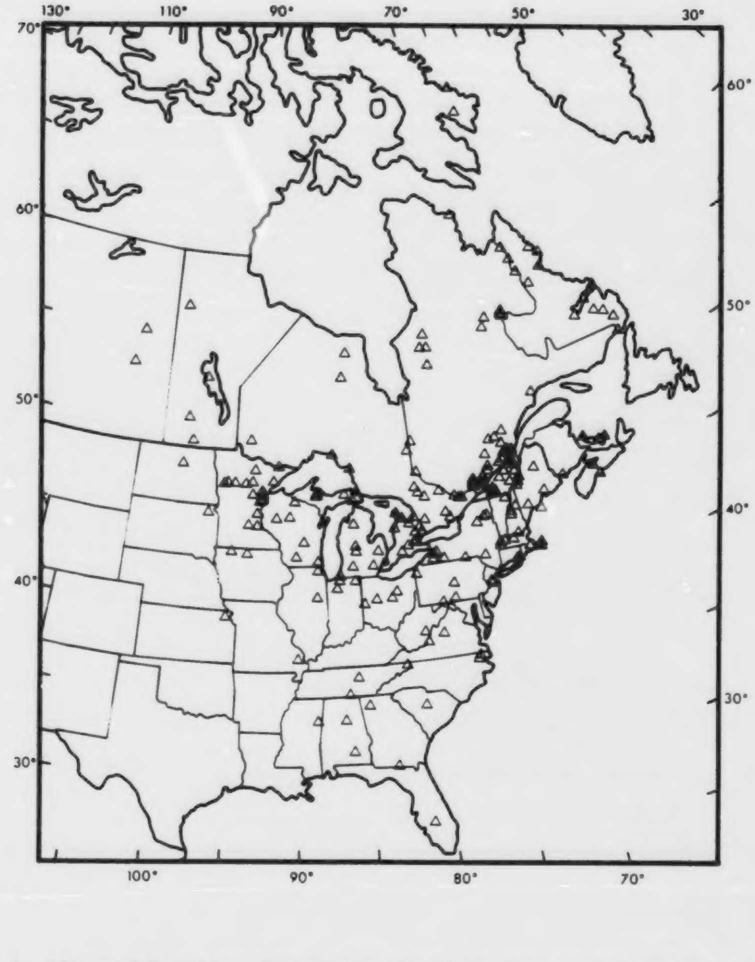

130°
110°
90°
70°
50°
30°
70°
60°
50°
40°
30°
60°
50°
40°
30°
100°
90°
80°
70°

FIGURE 2: *Sites with pollen data interpolated to 6000 yr B.P. Table 2 lists the site names and locations.*

(1) Selection of all sites with pollen data at 6000 yr B.P.

(2) Use of radiocarbon dates and age models to assign estimated ages to each depth with pollen counts at each of the sites with fossil data.

(3) Calculation of pollen percentages initially using a sum of all tree, shrub, and herb pollen and later using the same pollen sum as was used in developing the regression equation assigned to that sample.

(4) Estimation of the pollen percentages for 6000 yr B.P. by linear interpolation between the two nearest samples with ages bracketing 6000 yr B.P.

Bernabo and Webb (1977), Webb *et al.* (1983) and Webb *et al.* (1984) have presented maps showing how the percentages of pollen data for the major pollen types changed during the past 10,000 years. Data from Richard (1981), Lamb (1982), Short and Nichols (1977), and Andrews *et al.* (1980) have allowed the maps to be extended into northern Canada (Figure 2, Table 2).

CALIBRATION PROCEDURE

The steps used first to develop a regression equation and then to apply it to fossil data include: (1) the choice of the calibration region (the geographic area in which the modern pollen and climatic stations are located); (2) the selection of the pollen types used both in the pollen sum and as candidate independent variables in the regression equation; (3) the deletion of outliers and the transformation of certain pollen variables to gain linear relationships between the pollen and climatic variables; (4) the calculation of initial regression equations and checks of the statistical assumptions; (5) the calculation of the final multiple regression equations; and (6) the choice of the samples of fossil pollen to which the equation can be applied. Howe and Webb (1983) and Bartlein and Webb (1984B) have described steps (2) and (5) in detail, and Arigo *et al.* (1982) and Bartlein and Webb (1984A) described the SPSS (Nie *et al.* 1981) and BMDP (Dixon 1981) computer programs needed to complete the analyses required by these steps.

The large size of eastern North America in comparison with the distribution of each pollen type required the use of 13 different calibration regions. Although Howe and Webb (1983) and Bartlein and Webb (1984B) provided general guidelines for the choice of one calibration region, they did not discuss how best to choose the several regions needed to produce an isotherm map for the entire subcontinental area. We therefore tried to find methods to aid both the initial choice of calibration regions (Step 1) and the final choice of which equation to use at each site with pollen data from 6000 yr B.P. (Step 6).

Choice of the Calibration Region (Step 1)

According to Howe and Webb (1983), the bivariate relationship should be monotonic between the climatic variable being estimated and each of the major pollen types. Ditonic relationships, however, occur when the pollen data are plotted against latitude for a transect of sites across North America (Figure 1). For example, the percentages of oak *(Quercus)* pollen are negatively related to mean July temperature south of 37°N, and positively related north of there (Figure 3). Two different temperatures can therefore be associated with a given percentage of oak pollen. For other pollen types like birch *(Betula)* even more complicated relationships exist (Figure 3). As temperature steadily decreases toward the north over eastern North America, the percentages of *Betula* pollen first increase to a maximum around 45°N in the northern mixed forest, then decrease into the boreal forest, and finally rise again as shrub and dwarf birch populations increase at the northern edge of the boreal forest (Figure 3).

The strategy for dealing with such nonlinearities is to partition eastern North America into a number of overlapping subregions, within which the relationships between pollen and climate variables are clear and monotonic. This partitioning allows the highly nonlinear relationships to be approximated by relationships that are locally linear or, at least, monotonic. The selection of a subregion that is relatively small (perhaps 5° latitude by 5° longitude) may yield clear relationships, but will increase the risk that the resulting equations may produce an undesirable extrapolation when applied to fossil-pollen data. We have therefore attempted to keep the subregions relatively large (about 10° by 10°), and a reasonable operational rule seems to be that most regions should extend from the centre of one major vegetational type to the centre of another. We identified the initial boundaries between calibration regions by examining scatter diagrams of the abundances of the pollen types along longitudinal and latitudinal transects (Figure 3) and by noting the location of inflection points in the geographical trends of the pollen variables. Because temperature and precipitation in eastern North America vary with latitude and longitude, this procedure for assigning boundaries yielded pollen/climate relationships that were clear and monotonic.

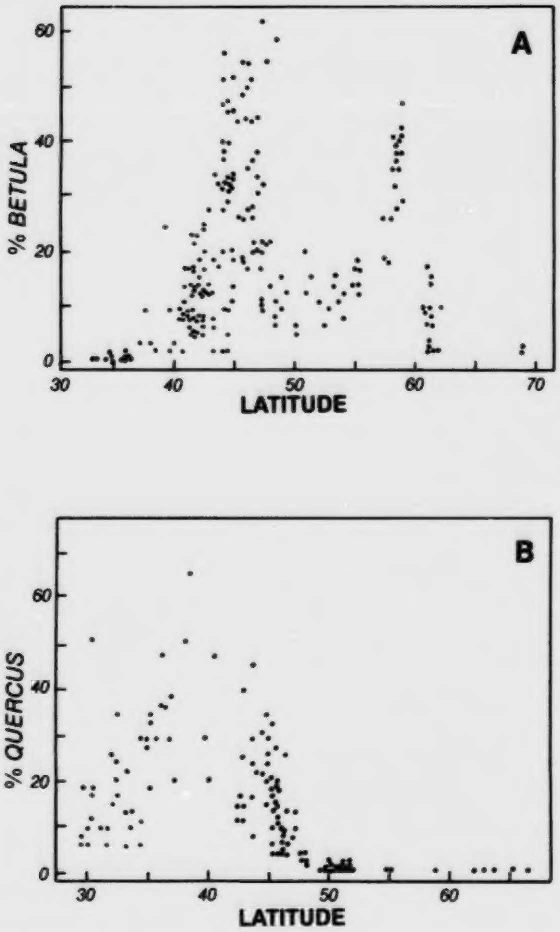


FIGURE 3: *Scatter diagrams of: (A) the percent birch (*Betula*); and (B) the percent oak (*Quercus*) pollen versus latitude.*

In the longitudinal band 90° to 95°N, for example, inflection points occurred at about 39° to 40°N (the latitudinal trends of *Quercus* and *Pinus* pollen reverse), around 45° to 46°N (the trends of *Betula* and *Acer* reverse), near 50°N (the trend for *Pinus* reverses again), and around 54° to 55°N (the trends for *Picea* and *Betula* reverse) (Figure 3). We located the boundaries for calibration regions at these latitudes and used scatter diagrams for other pollen types to locate the other boundaries. In this fashion, we identified 13 calibration regions that cover eastern North America (Figure 4).

Initial Selection of Pollen Types (Step 2)

Pollen types were chosen as candidate predictors in the regression equation by criteria for their numerical importance. Within a specific calibration region, a particular pollen type is considered to be important if: (a) its mean is greater than 1%; or (b) its maximum is greater than 5% (Howe and Webb 1983). Pollen types not meeting these criteria contribute mainly noise to any pollen/climate relationship (Webb and Clark 1977; Howe and Webb 1983). For the region from 45° to 55°N and 85° to 105°W, for example, eight pollen types met these criteria: *Picea*, *Betula*, *Pinus*, *Tsuga*, *Quercus*, *Alnus*, Gramineae; and a sum of prairie-forb pollen (*Artemisia* + Compositae + Chenopodiaceae/Amaranthaceae). The pollen percentages were then recalculated using a sum of just these pollen types.

Outlier Screening and Transformation of Certain Pollen Types (Step 3)

Once a calibration region was chosen, the relationships between a climate variable and the pollen variables might be monotonic but still not linear. Nonlinearity in the relationships between the dependent variable and the independent variable is a violation of one of the underlying assumptions of regression analysis. To linearize the relationships between a climate variable and several pollen predictor variables, we transformed each pollen variable separately because no single transformation of the climatic variable can linearize all of the bivariate relationships between it and the several pollen types. We chose transformations for each pollen type by inspecting bivariate scatter diagrams between the climate variable and each of the pollen types. These scatter diagrams also showed whether any outliers (i.e. unusual or anomalous observations) existed. Such observations can give a spurious impression of how a pollen type is related to the climatic variable, and

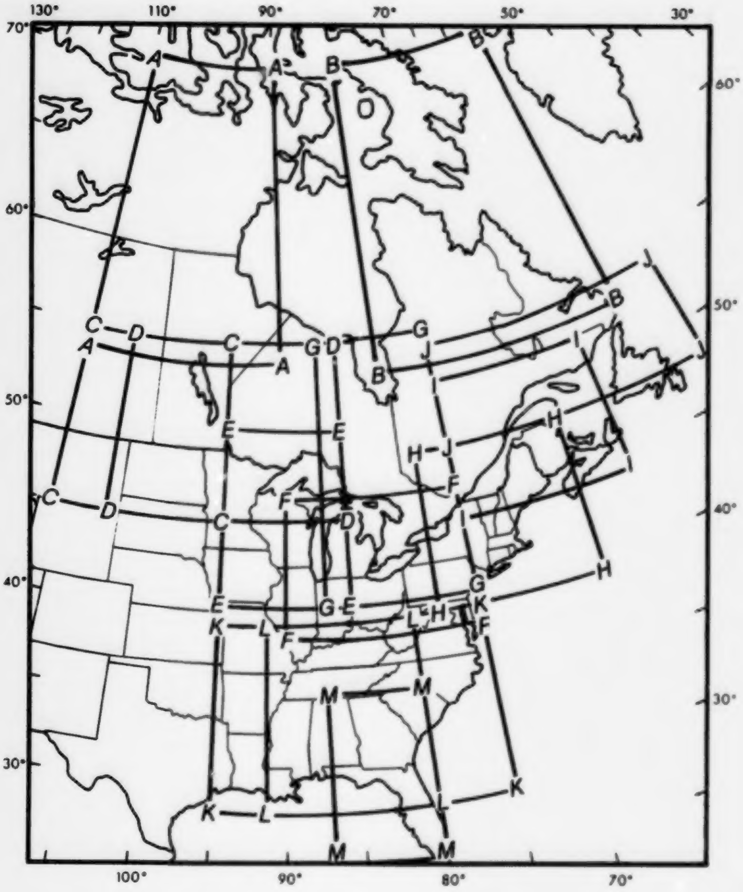

130°
110°
90°
70°
50°
30°
70°
60°
50°
40°
30°
60°
50°
40°
30°
100°
90°
80°
70°
A
A
B
B
D
C—D
A
C
A
G D
G
B
B
J
J
I
I
E
E
H—J
H
F
C—D
C
F
D
I
I
H
G
E
G E
L H
K
K—L
F
F
M
M
K
K—L
L
M
M

FIGURE 4: *The thirteen regions for which regression equations were obtained. The corner-point symbols in the figure correspond to the region designations in Table 3.*

may make the pollen type seem more important or less important than it really is. The choice of transformations and the screening for outliers allowed us to reduce the chances that model specification errors would occur.

For mean July temperature in the region 45° to 55°N, and 85° to 105°W, the relationship between temperature and spruce pollen was linearized by raising the spruce percentages to the 0.5 power (Figure 5), and that between oak pollen and temperature by raising the oak percentages to the 0.25 power. The scatter diagrams for oak pollen contained five observations with low oak percentages and high temperatures that plot well above the main scatter of points. These points are all located in the southwestern corner of this calibration region, and the low-oak values are more the result of dry conditions there than high temperatures. These observations diminish the importance of oak pollen as a predictor of temperature, and may therefore increase the chance that oak pollen would have an inappropriately low weighting in the regression equation or even be omitted from the final regression equation. Such anomalous observations that represent a different pollen/climate relationship from the rest of the data in the region are usually removed from the data set at this step in the analysis.

Initial Regressions (Step 4)

The first regression equation examined includes all pollen types in the pollen sum and provides both an asymptotic estimate of the "true" error variance and an initial display of several regression diagnostic statistics. These statistics may reveal further evidence for the existence of outliers or overly influential observations (Bartlein and Webb 1984B). Because the pollen variables are often highly correlated, the inclusion of redundant predictors in this "full" model will likely degrade the estimates of the regression coefficients. A subset of predictor variables is therefore needed. If a useful pollen type is omitted from the model, however, the estimates may again be degraded; and, if the spatial variations of the dependent climate variable are inadequately explained, the errors will not be independent, and the coefficients may be further degraded. The method we use to select independent variables in the final regression equation is designed to avoid these pitfalls. It requires an estimate of the true error variance, which is provided by this initial full regression model, and uses Mallow's Cp statistic to help select the "best" equation among the set of all possible regression equations (Bartlein and Webb 1984B).

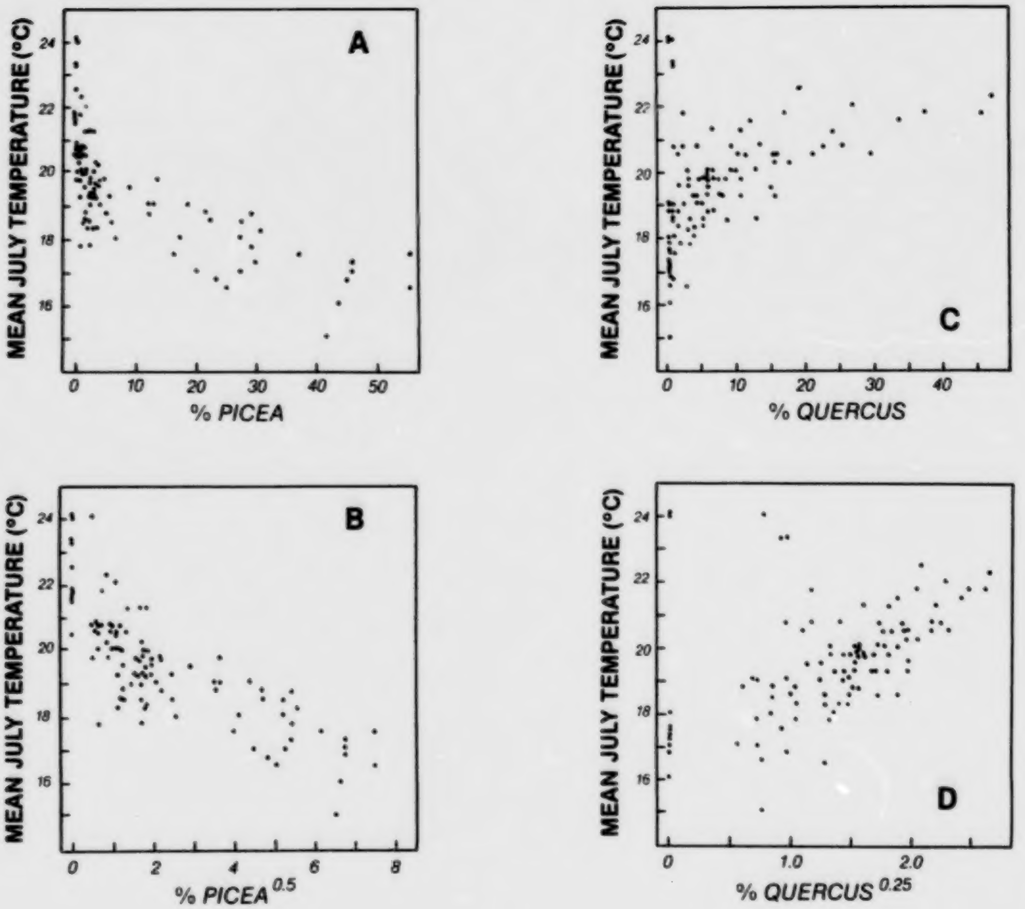


FIGURE 5: *Scatter diagrams of mean July temperature versus: (A) the percentage of spruce (*Picea*) pollen; (B) the percentage of spruce pollen raised to the 0.5 power; (C) the percentage of oak (*Quercus*) pollen; and (D) the percentage of oak pollen raised to the 0.25 power.*

Given the initial regression equation, multivariate outliers can be identified and deleted. These outliers are samples whose values for individual pollen types may be normal (i.e. within the range of that type for the calibration region), but which contain a combination of pollen values that are anomalous relative to the combinations of values in the rest of the observations. For instance, between 40°N and 50°N samples may contain 40% oak or 40% spruce pollen, but samples with both 40% oak and 40% spruce pollen would be highly unusual. The presence of such observations in the data set can again give a misleading impression of the importance of an individual pollen type as a predictor variable. The regression diagnostic statistics described in Bartlein and Webb (1984B) aid identification of these observations. These statistics also check the adequacy of the transformations chosen in the previous step. Further refinements of the regression equation are then possible. Usually the regression analysis must be repeated several times, because as one multivariate outlier is deleted, others may come to light.

Final Regressions and Equation Checking (Step 5)

Once the multivariate outliers and overly influential observations were omitted, a final subset of useful predictors was selected by the "best possible subsets" procedure that uses the Cp statistic as a selection criterion (Table 3[1]). We examined these final equations using the graphical diagnostic procedures described by Bartlein and Webb (1984B) and found no serious assumption violations.

Assumptions Checked by Steps 1 to 5

Decisions made during the particular sequence of steps described above minimize violations of the basic assumptions that underlie ordinary least-squares estimation in the following ways:

(1) The correct selection of the calibration region helps to eliminate specification error in the final model and ensures that linear (or

---

[1] Located at end of paper.

linearizable) relationships exist between the pollen and climate variables.

(2) The selection of the numerically most important pollen types to serve as candidate predictors helps to diminish the effects that errors in the predictor variables may have on the final regression equation.

(3) The transformation of individual pollen types also helps to reduce specification error by linearizing the monotonic relationship between the climatic variable and each pollen type. The removal of outliers helps by eliminating extraneous variability from bivariate pollen/climate relationships.

(4) The elimination of multivariate outliers further helps to diminish specification error, and also reduces the chances that errors may be non-normally distributed or have inhomogeneous variance.

(5) The selection of a final regression equation with only a subset of predictors included helps to reduce the deleterious effects of collinearity, diminishes specification error, and helps to eliminate dependencies among the residuals caused by imperfect explanation of the dependent climatic variable.

By following these five steps in the development of regression equations, we have found that we will obtain equations whose reliability (when applied to fossil-pollen samples) is not undermined by the existence of serious violations of the assumptions that underlie regression analysis.

Equation Selection for Climate Reconstruction (Step 6)

Given the 13 equations for estimating mean July temperature, we faced the problem of determining which equation to apply to each of the pollen samples from 6000 yr B.P. We selected the appropriate equation for each sample by indentifying the calibration region that: (1) contains modern pollen data that are analogous to the fossil sample; and (2) has an equation that does not produce an unwarranted extraplolation when applied to the fossil sample. The modern samples that were analogous to a particular fossil sample were determined by the calculation of a numerical measure of the dissimilarity between the fossil sample and individual modern samples (Overpeck *et al.* 1984). Further refinement in the equation selection was necessary because many of the calibration regions overlap, and because for many fossil samples modern analogues exist in several calibration regions. We therefore determined the extent to which a given equation, when applied to a particular fossil sample, produces an estimate which is an extrapolation - one based on a pollen spectrum that is relatively unusual when compared to the collection of samples used to develop that equation. The degree of extrapolation can be measured by the Mahalanobis distance between the fossil sample and the centroid of the modern samples from a particular calibration region (see Weisberg 1980). These distances were computed using only the pollen types appearing in a final equation, and employing the transformations chosen in Step 3 above. We chose the equation giving the smallest Mahalanobis distance. We also relied on the vegetational history in different regions, as gained from the maps in Bernabo and Webb (1977), or Webb *et al.* (1984) to provide additional guidance in choosing the appropriate equation for each fossil sample. In general, the equation from the region in which the fossil data were located was selected, but the latitudinal variation of vegetation during the Holocene meant that some fossil samples required equations developed for adjacent calibration regions (Figure 6).

ISOTHERM MAP FOR 6000 YR B.P.

The estimated mean July temperatures for 6000 yr B.P. (Figure 7) possess a south-to-north gradient from 27°C to 8°C just like those of today (Figure 8). The region with the steepest temperature gradient, however, was much further to the north at 6000 yr B.P. than it is today. This band, from 12°C to 21°C lay between the Canada - United States border and central Québec at 6000 yr B.P. (Figure 7), but today is between the 18°C and 22°C isotherms and is located south from the Canada-United States border to Pennsylvania and southern Michigan (Figure 8).

The estimated mean July temperatures for 6000 yr B.P. were higher that those today in a region from central Québec to the southern United States (Figure 9). This region of higher temperatures narrows to the west and extends across Manitoba into Saskatchewan. The temperature estimates were much lower than those observed today in northern Québec and Labrador. A small residual ice sheet still existed in the centre of this region, and may have contributed to the colder conditions that were estimated for the north.

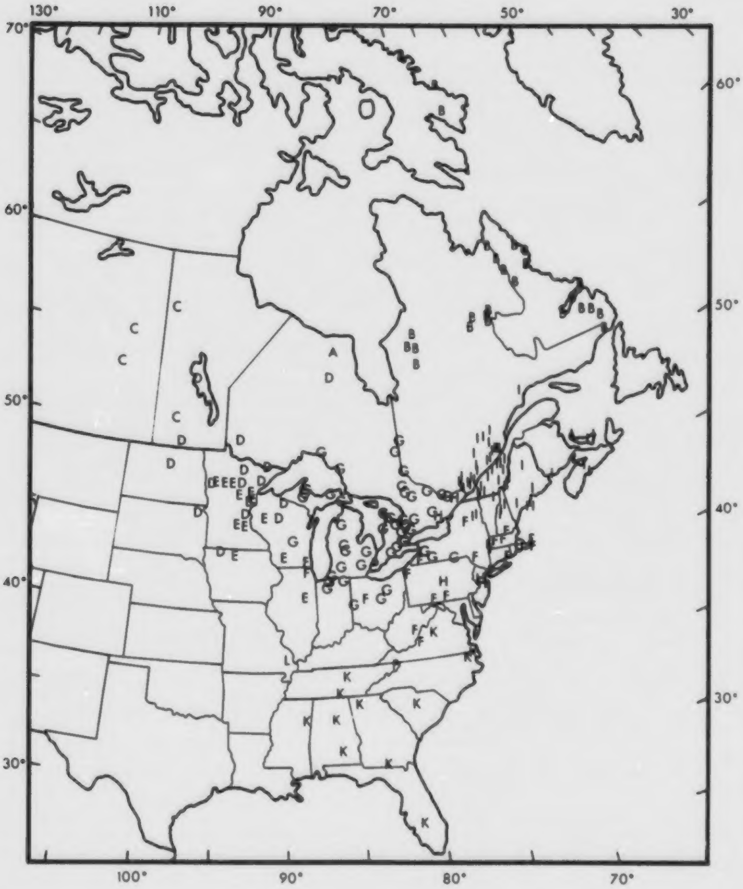

130°
110°
90°
70°
50°
30°
70°
60°
50°
40°
30°
60°
50°
40°
30°
100°
90°
80°
70°
A
B
C
D
E
F
G
H
I
K
L

FIGURE 6: *Regression equations used to reconstruct mean July temperature at 6000 yr B.P. Symbols correspond to region designations in Table 3.*

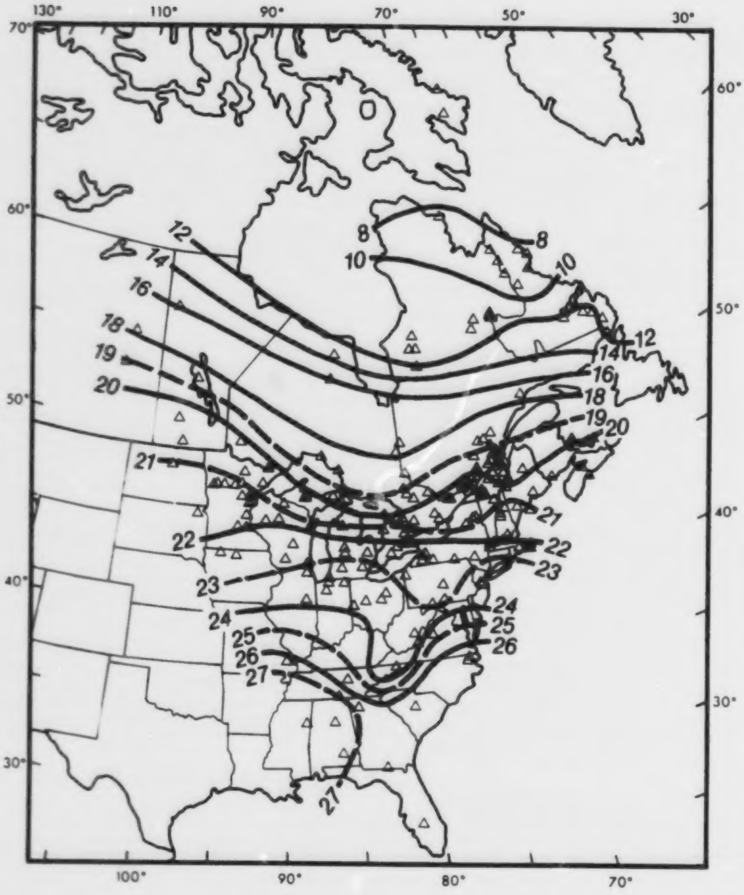

130°
110°
90°
70°
50°
30°
70°
60°
50°
40°
30°
60°
50°
40°
30°
100°
90°
80°
70°
8
8
10
10
12
12
14
14
16
16
18
18
19
19
20
20
21
21
22
22
23
23
24
24
25
25
26
26
27
27

FIGURE 7: *Isotherms for estimated mean July temperature (PC) at 6000 yr B.P.*

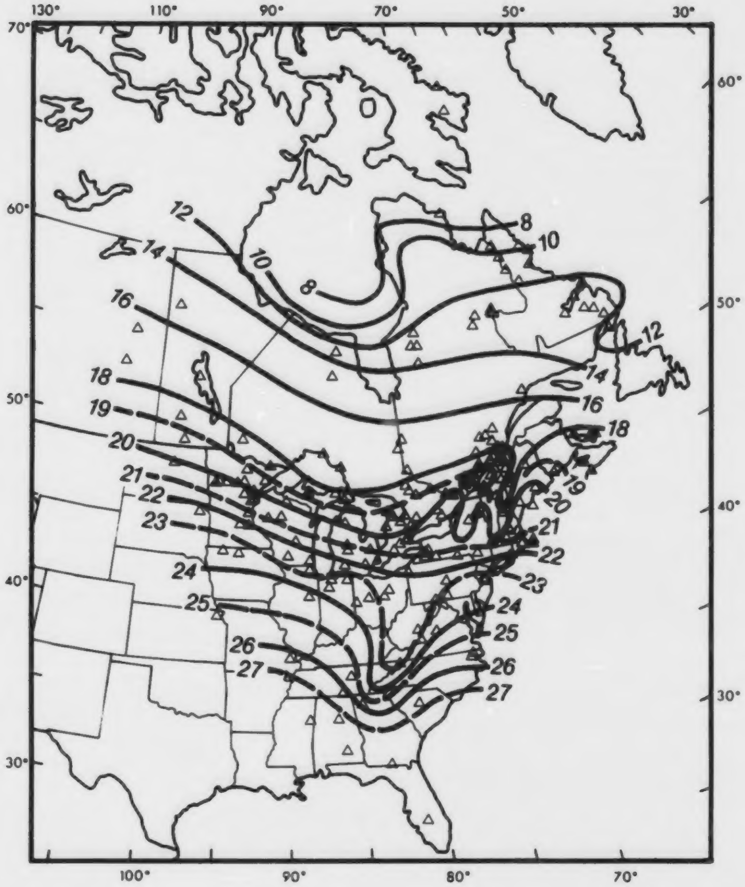

130°
110°
90°
70°
50°
30°
70°
60°
50°
40°
30°
60°
50°
40°
30°
100°
90°
80°
70°
12
14
16
18
19
20
21
22
23
24
25
26
27
10
8
8
10
12
14
16
18
19
20
21
22
23
24
25
26
27

FIGURE 8: *Isotherms for observed mean July temperature (PC) today.*

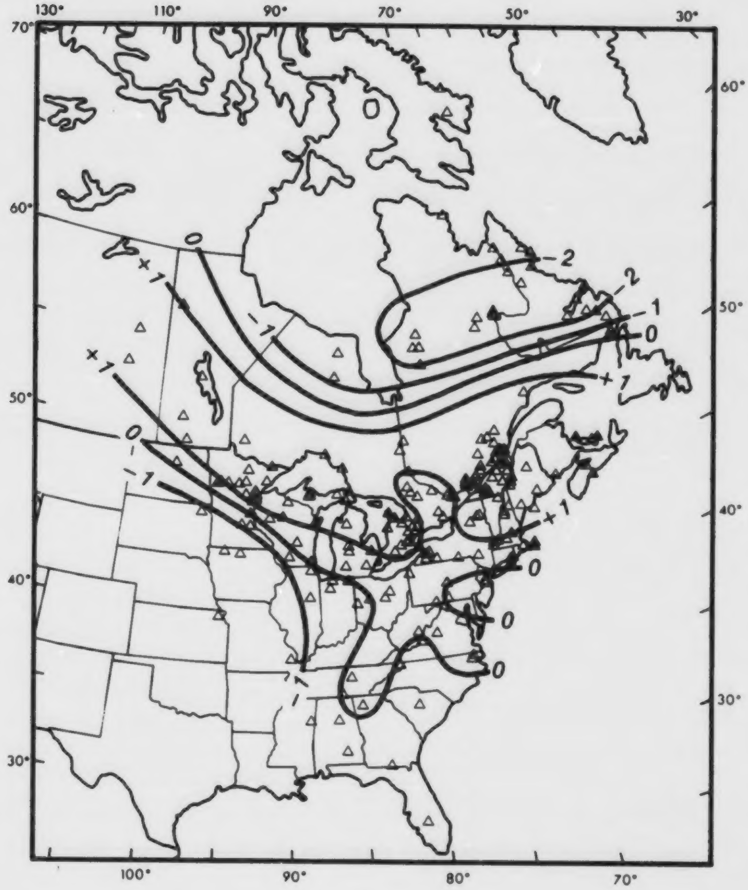

130°
110°
90°
70°
50°
30°
70°
60°
50°
40°
30°
60°
50°
40°
30°
100°
90°
80°
70°
0
+1
-1
+1
0
-1
-2
-2
-1
0
+1
+1
0
0
0
-1

FIGURE 9: *Difference map for mean July temperature (PC) between 6000 yr B.P. and today. Positive values indicate temperatures that were higher at 6000 yr B.P. than today.*

This pattern of temperature differences is broadly similar to the pattern generated with the NCAR (National Center for Atmospheric Research) global circulation model by Kutzbach and Guetter (1984A). They ran the model using the July solar radiation values for 6000 yr B.P., which were about 5.6% higher than today for the higher latitudes of the northern hemisphere. (This amplification of the seasonal cycle of solar radiation at 6000 yr B.P. was related to the different time of perihelion and different inclination of the earth's axis then.) Their experiment was similar to the work of Kutzbach (1981), Kutzbach and Otto-Bliesner (1982), and Kutzbach and Guetter (1984B) for 9000 yr B.P. In a subsequent paper we plan to describe the full evaluation of how well our reconstruction of Holocene temperature variations matched the model-simulated temperatures.

SUMMARY

Construction of an isotherm map for eastern North America at 6000 yr B.P. required a set of 13 multiple regression equations that estimate mean July temperature from a linear combination of pollen types. These equations were developed using over 1700 observations of modern pollen data, and then were applied to a network of 211 sites with pollen spectra dated to 6000 yr B.P. The isotherm map shows that mean July temperatures were higher at 6000 yr B.P. than today for a region from central Québec to the southern United States. This region of higher temperatures narrowed to the west and extended into Saskatchewan. The temperature estimates are much lower than those of today in northern Québec and Labrador, where a small residual ice sheet still existed at 6000 yr B.P.

ACKNOWLEDGEMENTS

United States National Science Foundation Grant (ATM81-11870) and a United States Department of Environment Contract (DEAC02-79EV10079) to Brown University supported most of the research in this article. The research was also supported by a United States National Science Foundation Grant (DEB81-15316) to Oak Ridge National Laboratory through a subcontract (19X-43321V) to the University of Oregon. We thank: R. Arigo, J. Avizinis, T. Benedict, A. Berman, C. Clifton, S. Klinkman, R. Mellor, C. Spitzer, S. Suter, and M. Symington for technical assistance.

REFERENCES

Andrews, J.T., W.M. Mode, and P.T. Davis. 1980. Holocene climate based on pollen transfer functions, eastern Canadian Arctic. Arctic and Alpine Research 12:41-64.

Arigo, R., S.E. Howe, and T. Webb, III. 1982. Computer programs for climatic calibration of pollen data. In: Paleohydrological Changes in the Temperate Zone in the Last 15,000 Years, IGCP 158B. Lake and Mire Environments, Project Guide Volume 3. Edited by: B.E. Berglund. Department of Quaternary Geology, University of Lund. pp. 79-109.

Bartlein, P.J., and T. Webb, III. 1984A. Annotated computer programs for climatic calibration of pollen data: a user's guide. In: Paleoecological Uses of Pollen Data. Edited by: J. Lentin. American Association of Stratigraphic Palynologists, Contributions. (In press).

_______. 1984B. Paleoclimatic interpretation of Holocene pollen data: statistical considerations. In: Paleoecological Uses of Pollen Data. Edited by: J. Lentin. American Association of Stratigraphic Palynologists, Contributions. (In press).

Bernabo, J.C., and T. Webb, III. 1977. Changing patterns in the Holocene pollen record from northeastern North America: a mapped summary. Quaternary Research 8:64-96.

Birks, H.J.B., T. Webb, III, and A. Berti. 1975. Numerical analysis of pollen samples from central Canada: a comparison of methods. Review of Paleobotany and Palynology 20:133-169.

Butzer, K.W. 1980. Adaptation to global environmental change. Professional Geographer 32:269-278.

CLIMAP Project Members. 1976. The surface of the ice-age earth. Science 191:1131-1136.

Cole, H.S. 1969. Objective reconstruction of the paleoclimatic record through the application of eigenvectors of present-day pollen spectra and climate to the late-Quaternary pollen stratigraphy. Ph.D. thesis, University of Wisconsin-Madison. 110 pp.

Delcourt, P.A., H.R. Delcourt, and T. Webb, III. 1984. Atlas of paired isophyte and isopoll maps for important eastern North American taxa. American Association of Stratigraphic Palynologists, Contributions. (In press).

Dixon, W.J. (General editor). 1981. BMDP statistical software. University of California Press, Berkeley. 726 pp.

Heusser, C.J., and S.S. Streeter. 1980. A temperature and precipitation record of the past 16,000 years in southern Chile. Science 210:1345-1347.

Howe, S.E., and T. Webb, III. 1977. Testing the statistical assumptions of paleoclimatic calibration functions. Preprint Volume: Fifth Conference of Probability and Statistics in the Atmospheric Sciences. American Meteorological Society, Boston. pp. 152-157.

_______. 1983. Calibrating pollen data in climatic terms: improving the methods. Quaternary Science Reviews 2:17-51.

Kay, P.A. 1979. Multivariate statistical estimates of Holocene vegetation and climatic change, forest-tundra transition zone, N.W.T., Canada. Quaternary Research 11:125-140.

Kellogg, W.W. 1978. Global influences of mankind on climate. In: Climatic Change. Edited by: J. Gribben. Cambridge University Press, Cambridge. pp. 205-227.

Kutzbach, J.E. 1981. Monsoon climate of the early Holocene: climate experiment with the earth's orbital parameters for 9000 years ago. Science 214:59-61.

Kutzbach, J.E., and P.J. Guetter. 1984A. Sensitivity of late-glacial and Holocene climates to the combined effects of orbital parameter changes and lower boundary condition changes: snapshot simulations with a general circulation model for 18,000, 9000, and 6000 years ago. Journal of Glaciology. (In press).

______. 1984B. The sensitivity of monsoon climates to orbital parameter changes for 9000 yr B.P.: experiments with the NCAR general circulation model. In: Milankovitch and Climatic Change. Edited by: J. Imbrie and A. Berger. Elsevier Scientific Publishing Co., Amsterdam. (In press).

Kutzbach J.E., and B. Otto-Bliesner. 1982. The sensitivity of the African-Asian monsoonal climate to orbital parameter changes for 9000 years B.P. in a low-resolution general circulation model. Journal of the Atmospheric Sciences 39:1177-1188.

Lamb, H.F. 1982. Late Quaternary vegetational history of the forest-tundra ecotone in north-central Labrador. Ph.D. thesis, University of Cambridge. 195 pp.

Nie, N.H., C.R. Hull, J.G. Jenkins, K. Steinbrenner, and D.H. Bent. 1975. SPSS: Statistical Package for the Social Sciences. 2nd Edition. McGraw-Hill, New York. 675 pp.

Overpeck, J.T., T. Webb, III, and I.C. Prentice. 1984. Quantitative interpretation of fossil pollen spectra: dissimilarity coefficients and the method of modern analogs. Quaternary Research. (Submitted for publication).

Peterson, G.M., T. Webb, III, J.E. Kutzbach, T. Van der Hammen, T.J. Wijmstra, and F.A. Street. 1979. The continental record of environmental conditions at 18,000 yr B.P.: an initial evaluation. Quaternary Research 12:47-82.

Richard, P.J.H. 1981. Paleophytogeographie postglaciaire en Ungava par l'analyse pollinique. Paleo-Québec 13:1-153.

Short, S.K., and H. Nichols. 1977. Holocene pollen diagrams from subarctic Labrador-Ungava: vegetational history and climatic change. Arctic and Alpine Research 9:265-290.

Street, F.A., and A.T. Grove. 1979. Global maps of lake-level fluctuations since 30,000 yr B.P. Quaternary Research 12:83-118.

Webb, T., III. 1984. Pollen analysis and paleoclimatology. In: Paleoclimate Data and Modeling. Edited by: A.D. Hecht. J. Wiley and Sons, Inc., New York (In press).

Webb, T., III, and R.A. Bryson. 1972. Late-and postglacial climatic change in the northern Midwest, U.S.A.: quantitative estimates derived from fossil pollen spectra. Quaternary Research 2:70-115.

Webb, T., III and D.R. Clark. 1977. Calibrating micropaleontological data in climatic terms: a critical review. Annals of the New York Academy of Sciences 228:93-118.

Webb, T., III, and J.H. McAndrews. 1976. Corresponding patterns of contemporary pollen and vegetation in central North America. Geological Society of America Memoir 145:267-299.

Webb, T., III, P.J.H. Richard, and R.J. Mott. 1983. A mapped history of Holocene vegetation in southern Québec. In: Climatic Change in Canada 3. Edited by: C.R. Harington. Syllogeus No. 49:273-336.

Webb, T., III, E.J. Cushing, and H.E. Wright, Jr. 1984. Holocene changes in the vegetation of the Midwest. In: Late-Quaternary Environments of the United States, Volume 2, The Holocene. Edited by: H.E. Wright, Jr. University of Minnesota Press, Minneapolis. pp. 142-165.

Weisberg, S. 1980. Applied linear regression. Wiley, New York. 283 pp.

REFERENCES FOR 6000 YR B.P. DATA

Anderson, T.W. 1980. Holocene vegetation and climatic history of Prince Edward Island, Canada. Canadian Journal of Earth Sciences 17:1152-1165.

Bailey, R.E. 1972. Late- and postglacial environmental changes in northwestern Indiana. Ph.D. thesis, Indiana University, Bloomington. 72 pp.

Bailey, R.E., and P.J. Ahearn. 1981. A late- and postglacial pollen record from Chippewa Bog, Lapeer Co., MI: further examination of white pine and beech immigration into the central Great Lakes region. In: Geobotany II. Edited by: R.C. Romans. Plenum Press, New York. pp. 53-74.

Baker, R.G. 1970. A radiocarbon-dated pollen chronology for Wisconsin: Disterhaft Farm Bog revisited. Geological Society of America Abstracts 2:488.

Barclay, F.H. 1957. The natural vegetation of Johnson County, Tennessee past and present. Ph.D. thesis, University of Tennessee, Knoxville. 147 pp.

Bostwick, L.K. 1978. An environmental framework for cultural change in Maine: pollen influx and percentage diagrams from Monhegan Island. M.S. thesis, University of Maine-Orono. 68 pp.

Brubaker, L.B. 1975. Postglacial forest patterns associated with till and outwash in northcentral upper Michigan. Quaternary Research 5:499-527.

Comtois, P. 1982. Histoire holocène du climat et de la végétation à Lanoraie (Québec). Canadian Journal of Earth Sciences 19:1938-1952.

Craig, A.J. 1969. Vegetational history of the Shenandoah Valley, Virginia. Geological Society of America Special Paper 123:283-296.

______. 1972. Pollen influx to laminated sediments: a pollen diagram from northeastern Minnesota. Ecology 53:46-57.

Davis, A.M. 1977. The prairie-deciduous forest ecotone in the upper Middle West. Annals of the American Association of American Geographers 67:204-213.

Davis, M.B. 1969. Climatic changes in southern Connecticut recorded by pollen deposition at Rogers Lake. Ecology 50:409-422.

Davis, P.T. 1980. Late Holocene glacial, vegetational, and climatic history of Pangnirtung and Kingnait Fiord area, Baffin Island, Canada. Ph.D. thesis, University of Colorado, Boulder. 366 pp.

Davis, R.B., T.E. Bradstreet, R.Stackenrath, Jr., and H.W. Borns, Jr. 1975. Vegetation and associated environments during the past 14,000 years near Moulton Pond, Maine. Quaternary Research 5:435-465.

Delcourt, H.R. 1979. Late Quaternary vegetation history of the eastern Highland Rim and adjacent Cumberland Plateau of Tennessee. Ecological Monographs 49:255-280.

Delcourt, H.R., P.A. Delcourt, and E.C. Spiker. 1983. A 12 000-year record of forest history from Cahaba Pond, St. Clair County, Alabama. Ecology 64:871-887.

Delcourt, P.A. 1980. Goshen Springs: late Quaternary vegetation record for southern Alabama. Ecology 61:371-386.

Durkee, L.H. 1971. A pollen profile from Woden Bog, Hancock County, Iowa. Ecology 52:835-844.

Fries, M. 1962. Pollen profiles of late Pleistocene and recent sediments at Weber Lake, northeastern Minnesota. Ecology 43:295-308.

Futyma, R.P. 1982. Postglacial vegetation of eastern upper Michigan. Ph.D. thesis, University of Michigan, Ann Arbor. 426 pp.

Gauthier, R. 1981. Histoire de la colonisation vegetale postglaciaire des Montérégiennes: deux sites du mont Saint-Bruno. Mémoire M.Sc., Université de Montréal. 107 pp.

Gilliam, J.A., R.O. Kapp, and R.D. Bogue. 1967. A post-Wisconsin pollen sequence from Vestaberg Bog, Moncalm County, Michigan. Michigan Academy of Science, Arts and Letters 52:3-17.

Grüger, J. 1973. Studies on the late Quaternary vegetation history of northeastern Kansas. Geological Society of America Bulletin 84:239-250.

Hadden, K.A. 1975. A pollen diagram from a postglacial peat bog in Hants County, Nova Scotia. Canadian Journal of Botany 53:39-47.

Heide, K.M. 1981. Late Quaternary vegetational history of northcentral Wisconsin, U.S.A.: estimating forest composition from pollen data. Ph.D. thesis, Brown University, Providence. 312 pp.

Jacobson, G.L. 1979. The paleoecology of white pine in Minnesota. Journal of Ecology 67:669-728.

Janssen, C.R. 1968. Myrtle Lake: a late- and postglacial pollen diagram from northern Minnesota. Canadian Journal of Botany 46:1397-1410.

Jordan, R.H. 1975. Pollen diagrams from Hamilton Inlet, central Labrador, and their environmental implications for the northern maritime Arctic. Arctic Anthropology 12:92-116.

Kapp, R.O. 1977. Paleoecology in central lower Michigan, field trip guide. Paleoecology Section, Ecological Society of America.

Karrow, P.F., T.W. Anderson, A.H. Clarke, L.D. Delorme, and M.R. Sreenivasa. 1975. Stratigraphy, paleontology, and age of Lake Algonquin sediments in southwestern Ontario, Canada. Quaternary Research 5:49-87.

Kerfoot, W.C. 1974. Net accumulation rates and the history of cladoceran communities. Ecology 55:51-61.

King, J.E. 1981. Late Quaternary vegetational history of Illinois. Ecological Monographs 51:43-62.

King, J.E., and W.H. Allen, Jr. 1979. A Holocene vegetation record from the Mississippi River valley, southeastern Missouri. Quaternary Research 8:307-323.

Labelle, C., and P.J.H. Richard. 1981. Végétation tardiglaciaire et postglaciare au sud-est du Parc des Laurentides, Québec. Géographie physique et Quaternaire 35:345-359.

Lamb, H.F. 1978. Post-glacial vegetation change in southern Labrador. M.S. thesis, University of Minnesota, Minneapolis. 101 pp.

______. 1980. Late Quaternary vegetational history of southeastern Labrador. Arctic and Alpine Research 12:117-135.

Lawrenz, R. 1975. Biostratigraphic study of Green Lake Michigan. M.S. thesis, Central Michigan University, Mt. Pleasant.

Livingstone, D.A. 1968. Some interstadial and postglacial pollen diagrams from eastern Canada. Ecological Monographs 38:87-125.

Manny, B.A., R.G. Wetzel, and R.E. Bailey. 1978. Paleolimnological sedimentation of organic carbon, nitrogen, phosphorous, fossil pigments, pollen and diatoms in a hypereutrophic hardwater lake: a case history for eutrophication. Polskie Archiwum Hydrobiologii 25:243-267.

McAndrews, J.H. 1966. Postglacial history of prairie, savanna and forest in northeastern Minnesota. Torrey Botanical Club Memoirs 22:1-72.

______. 1970. Fossil pollen and our changing landscape and climate. Rotunda 3:30-37.

______. 1981. Late Quaternary climate of Ontario: temperature trends from the fossil pollen record. In: Quaternary Paleoclimate. Edited by: W.C. Mahaney. Geoabstracts, Norwich, England. pp. 319-333.

______. 1982. Holocene environment of a fossil bison from Kenora, Ontario. Ontario Archaeology 37:41-51.

McAndrews, J.H., J.L. Riley, and A.M. Davis. 1982. Vegetation history of the Hudson Bay Lowland: a postglacial pollen diagram from Sutton Ridge. Naturaliste canadien 109:597-608.

McDowell, L.L., R.M. Dole, Jr., M. Howard, and R.A. Farrington. 1971. Palynology and radiocarbon chronology of Bugbee Wildflower Sanctuary and Natural Area, Caledonia County, Vermont. Pollen et Spores 13:73-91.

Miller, N.G. 1973. Late-glacial and postglacial vegetation change in southwestern New York State. New York State Museum and Science Service Bulletin 420:1-102.

Mode, W. 1980. Quaternary stratigraphy and palynology of the Clyde Foreland, Baffin Island, N.W.T., Canada. Ph.D. thesis, University of Colorado, Boulder. 219 pp.

Mott, R.J. 1973. Palynological studies in central Saskatchewan, pollen stratigraphy from lake sediment sequences. Geological Survey of Canada Paper 72-49:1-18.

______. 1975. Palynological studies of lake sediment profiles from southwestern New Brunswick. Canadian Journal of Earth Sciences 12:273-288.

______. 1976. A Holocene pollen profile from the Sept-Iles area, Québec. Naturaliste canadien 103:457-467.

______. 1977. Late Pleistocene and Holocene palynology in southeastern Québec. Géographie physique et Quaternaire 31:139-149.

Mott, R.J., and M. Camfield. 1969. Palynological studies in the Ottawa area. Geological Survey of Canada Paper 69-38:1-16.

Mott, R.J., and L.D. Farley-Gill. 1978. A late-Quaternary pollen profile from Woodstock, Ontario. Canadian Journal of Earth Sciences 15:1101-1111.

______. 1981. Two late Quaternary pollen profiles from Gatineau Park, Quebec. Geological Survey of Canada Paper 80-31:1-10.

Nicholas, J. 1968. Late Pleistocene palynology of southeastern New York and northern New Jersey. Ph.D. thesis, New York University.

Nichols, H. 1967. Central Canadian palynology and its relevance to northwestern Europe in the late Quaternary period. Review of Palaeobotany and Palynology 2:231-243.

Ogden, J.G., III. 1966. Forest history of Ohio: radiocarbon dates and pollen stratigraphy of Silver Lake, Ohio. Ohio Journal of Science 66:387-400.

Ogden, J.G., III, and R.J. Hay. 1967. Ohio Wesleyan University natural radiocarbon measurements III. Radiocarbon 9:316-332.

Overpeck, J.T. 1984. A palynological and sedimentological investigation of a late Quaternary peat bog in the south-central Adirondack Mountains, New York. Geological Society of America Bulletin. (Submitted for publication).

Peters, A., and T. Webb, III. 1979. A radiocarbon-dated pollen diagram from west-central Wisconsin. Bulletin of the Ecological Society of America 60:102.

Richard, P. 1977. Histoire post-wisconsinienne de la végétation du Québec méridional par l'analyse pollinique. Service de la Recherche, Direction Général des Forêts, Ministère des Terres et Forêts, Québec. tome 1, 312 pp. tome 2, 142 pp.

______. 1979. Contribution à l'histoire postglaciaire de la végétation au nord-est de la Jamésie, Nouveau-Québec. Géographie physique et Quaternaire 33:93-112.

Richard, P., A. Larouche, and M. Bouchard. 1982. Age de la déglaciation finale et histoire postglaciaire de la végétation dans la partie centrale du Nouveau-Québec. Géographie physique et Quaternaire 36:63-90.

Ritchie, J.C. 1969. Absolute pollen frequencies and carbon-14 age of a section of Holocene lake sediment from the Riding Mountain area of Manitoba. Canadian Journal of Botany 47:1345-1349.

Ritchie, J.C., and S. Lichti-Federovich. 1968. Holocene pollen assemblages from the Tiger Hills, Manitoba. Canadian Journal of Earth Sciences 5:873-880.

Ritchie, J.C., and K.A. Hadden. 1975. Pollen stratigraphy of Holocene sediments from the Grand Rapids area, Manitoba, Canada. Review of Palaeobotany and Palynology 19:193-202.

Saarnisto, M. 1974. The deglaciation history of the Lake Superior region and its climatic implications. Quaternary Research 4:316-339.

______. 1975. Stratigraphic studies on the shoreline displacement of Lake Superior. Canadian Journal of Earth Sciences 12:300-319.

Savoie, L., and P. Richard. 1979. Paléophytogéographie de l'episode de Saint Narcisse dans la région de Sainte-Agathe, Québec. Géographie physique et Quaternaire 33:175-188.

Shane, L.C.K. 1976. Late-glacial and postglacial palynology and chronology of Drake County, west-central Ohio. Ph.D. thesis, Kent State University, Kent.

Sheehan, M.C., and D.R. Whitehead. 1981. Holocene vegetational changes in the Tombigbee River Valley, eastern Mississippi. Bulletin of the Ecological Society of America 62:126.

Short, S.K., and H. Nichols. 1977. Holocene pollen diagrams from subarctic Labrador-Ungava: vegetational history and climatic change. Arctic and Alpine Research 9:265-290.

Spear, R.W. 1981. The history of high-elevation vegetation in the White Mountains of New Hampshire. Ph.D. thesis, University of Minnesota, Minneapolis. 215 pp.

Spear, R.W., and N.G. Miller. 1976. A radiocarbon-dated pollen diagram from the Allegheny Plateau of New York State. Journal of the Arnold Arboretum 57:369-403.

Stravers, L.K.S. 1981. Palynology and deglaciation history of the central Labrador-Ungava Peninsula. M.S. thesis, University of Colorado, Boulder. 171 pp.

Swain, P.C. 1979. The development of some bogs in eastern Minnesota. Ph.D.thesis. University of Minnesota, Minneapolis.

Terasmae, J. 1968. A discussion of deglaciation and the boreal forest in the northern Great Lakes region. Proceedings of the Entomological Society of Ontario 99:31-43.

Terasmae, J., and T.W. Anderson. 1970. Hypsithermal range extension of white pine (*Pinus strobus* L.) in Quebec, Canada. Canadian Journal of Earth Sciences 7:406-413.

Van Zant, K.L. 1979. Late-glacial and postglacial pollen and plant macrofossils from Lake West Okoboji, northwestern Iowa. Quaternary Research 12:358-380.

Vincent, J.S. 1973. A palynological study for the Little Clay Belt, northwestern Quebéc. Naturaliste canadien 100:59-69.

Waddington, J.C.B. 1969. A stratigraphic record of pollen influx to a lake in the Big Woods of Minnesota. Geological Society of America Special Paper 123:263-282.

Walker, P.C., and R.I. Hartman. 1960. The forest sequence of the Hartstown Bog area in western Pennsylvania. Ecology 41:461-474.

Watts, W.A. 1970. The full-glacial vegetation of northwestern Georgia. Ecology 51:17-33.

______. 1971. Postglacial and interglacial vegetation history of southern Georgia and central Florida. Ecology 52:676-690.

______. 1975. A late Quaternary record of vegetation from Lake Annie, south-central Florida. Geology 3:344-346.

______. 1979. Late Quaternary vegetation of central Appalachia and the New Jersey coastal plain. Ecological Monographs 49:427-469.

______. 1980. Late-Quaternary vegetation history at White Pond on the inner Coastal Plain of South Carolina. Quaternary Research 13:187-199.

Watts, W.A., and R.C. Bright. 1968. Pollen, seed, and mollusk analysis of a sediment core from Pickerel Lake, northeastern South Dakota. Geological Society of America Bulletin 79:855-876.

Webb, T., III. 1974. A vegetational history from northern Wisconsin. Evidence from modern and fossil pollen. American Midland Naturalist 92:12-32.

Webb, T., III, and R.A. Bryson. 1972. Late- and postglacial climatic change in the northern Midwest, U.S.A.: quantitative estimates derived from fossil pollen spectra by multivariate statistical analysis. Quaternary Research 2:70-115.

Whitehead, D.R. 1972. Developmental and environmental history of the Dismal Swamp. Ecological Monographs 42:301-315.

______. 1979. Late-glacial and post-glacial vegetational history of the Berkshires, western Massachusetts. Quaternary Research 12:333-357.

______. 1981. Late-Pleistocene vegetational changes in northeastern North Carolina. Ecological Monographs 51:451-471.

Whitehead, D.R., and T.L. Crisman. 1978. Paleolimnological studies of small New England (U.S.A.) ponds. Part I. Late-glacial and postglacial trophic oscillations. Polskie Archiwum Hydrobiologii 25:471-481.

Williams, A.S. 1974. Late-glacial--postglacial vegetational history of the Pretty Lake region, northeastern Indiana. United States Geological Survey Professional Paper 686:1-23.

Wright, H.E., Jr., and W.A. Watts. 1969. Glacial and vegetational history of northeastern Minnesota. Minnesota Geological Survey SP-11:1-59.

Wright, H.E., Jr., T.C. Winter, and H.L. Patton. 1963. Two pollen diagrams from southeastern Minnesota: problems in the late- and postglacial vegetational history. Geological Society of America Bulletin 74:1371-1396.

TABLE 1: *ASSUMPTIONS REQUIRED BY ORDINARY LEAST SQUARES REGRESSION ANALYSIS, AND THE EFFECTS, CAUSES, AND SYMPTOMS OF VIOLATING THESE ASSUMPTIONS (BARTLEIN AND WEBB 1984B).*

ASSUMPTION VIOLATED	EFFECTS ON OLS ESTIMATORS	MOST LIKELY CAUSES	SYMPTOMS AND TESTS	SOLUTIONS
1A: $E[e_i] = \sigma^2$	- $\underline{b}$ unbiased, but inefficient	- curvilinear relationship between $\underline{y}$ and $\underline{X}$ - calibration region too large - outliers	- patterns in residual plots - Goldfeld - Quandt test - residual maps	- transform $\underline{x}_j$'s - delete outliers - limit region
1B: $e_i \sim N(0, \sigma^2)$	- F and t-tests inappropriate - interval estimation of $\underline{y}$ inappropriate	- outliers - dependent variable highly non-normal	- nonlinear normal probability plot	- delete outliers - transform $\underline{y}$
1C: $E[e_i e_j] = 0$ all $i \neq j$	- $\underline{b}$ unbiased, but inefficient - F and t-tests misleading - σ^2 and R^2 inflated	- misspecification: a. omitted $\underline{x}_j$'s b. nonlinear relationships - outliers - calibration region too large	- pattern on residual maps - Moran test for spatial autocorrelation	- respecify model - delete outliers - limit region - use alternative estimation methods
2: $\underline{X}$ known without error	- $\underline{b}$ biased and inconsistent	- errors in sampling predictor variables	- select observations to minimize σ^2_X	- combine samples - use alternative estimation methods
3: $\underline{X}$ has rank p, where $p < n$	- large variances of parameter estimates - parameter estimates sensitive to observation deletion	- closed data set - common spatial trends in predictors - calibration region too small	- condition of $\underline{X}$ - correlations among parameters - correlations among predictors - tolerance values	- omit redundant predictors - principal component analysis of predictors
4: Correct model specification	- $\underline{b}$ biased and inconsistent	- omission of useful predictors - nonlinear relationships - calibration region too large - outliers	- spatial autocorrelation of residuals - many influential observations - patterns on residual maps	- respecify model - C_p selection criterion - delete outliers - limit region

TABLE 2: *SITES WITH POLLEN DATA FOR 6000 YR B.P. IN EASTERN NORTH AMERICA.*

Site Name	State/ Province	Latitude (degrees) (minutes)	Longitude (degrees) (minutes)	Elevation (m)	Number of C-14 Dates	Dating Quality	Site Type	Source
Albion	PQ	45 40 N	71 19 W	320	4	1	Lake	Richard, P., 1977
Alderdale	ON	46 3 N	79 12 W	240	1	4	Bog	Terasmae, J., 1968
Alexander Lake	LB	53 20 N	60 35 W	143	3	1	Lake	Jordan, R.H., 1975
Alexis Lake	LB	52 31 N	57 2 W	200	4	2	Lake	Lamb, H.F., 1978
Alfies Lake	ON	47 53 N	84 52 W	288	3	4	Lake	Saarnisto, M., 1974
Aliuk Pond	LB	54 35 N	57 22 W	25	2	2	Lake	Jordan, R.H., 1975
Allenberg Bog	NY	42 15 N	78 52 W	494	0	7	Bog	Miller, N.G., 1973
Anderson Pond	TN	36 2 N	85 30 W	305	10	2	Lake	Delcourt, H.R., 1979
Attawapisket	ON	53 0 N	85 10 W	100	1	2	Bog	Terasmae, J., 1968
Baie St. Paul -- Ange	PQ	47 28 N	70 41 W	640	7	1	Lake	Labelle, C. & P.Richard, 1981
Ballycroy Bog	ON	43 57 N	79 52 W	297	2	4	Bog	Karrow, P. et al., 1975
Base de Plein Air de St.Foy	PQ	46 47 N	71 20 W	16	2	2	Bog	Larouche, A., P.Richard, unpubl.
Basswood Road Lake	NB	45 15 N	67 20 W	106	3	1	Lake	Mott, R.J., 1975
Belmont Bog	NY	42 15 N	77 55 W	497	6	4	Bog	Spear, R. & N.G.Miller, 1976
Bereziuk	PQ	54 3 N	76 7 W	205	3	1	Lake	Richard, P., 1979
Berry Pond	MA	42 30 N	73 19 W	600	16	2	Lake	Whitehead, D.R., 1979
Big Pond	PA	39 46 N	78 33 W	634	1	5	Lake	Watts, W.A., 1979
Blackington Lake	ON	47 54 N	84 52 W	261	2	4	Lake	Saarnisto, M., 1975
Bog D	MN	47 11 N	95 10 W	457	4	3	Lake	McAndrews, J.H., 1966
Boundary	ME	45 34 N	70 41 W	603	5	3	Lake	Mott, R.J., 1977
Boundary Lake	PQ	55 15 N	67 24 W	525	6	1	Lake	Stravers, L.K., 1981
Brandreth Lake	NY	43 55 N	74 41 W	583	5	1	Bog	Overpeck, J.T., 1984
Brown's Lake Bog	OH	40 41 N	82 3 W	290	11	2	Bog	Ogden, J.G. III, unpubl.
Bugbee Bog	VT	44 22 N	72 9 W	398	4	2	Bog	McDowell, L.L. et al., 1971
BL-Tombigbee	AL	33 33 N	88 28 W	49	5	3	Lake	Sheehan, M.C. & D.R.Whitehead,1981
Cahaba Pond	AL	33 34 N	86 31 W	204	13	2	Lake	Delcourt, H.R. et al., 1983

TABLE 2: *SITES WITH POLLEN DATA FOR 6000 YR B.P. IN EASTERN NORTH AMERICA.* (Cont'd)

Site Name	State/ Province	Latitude (degrees) (minutes)	Longitude (degrees) (minutes)	Elevation (m)	Number of C-14 Dates	Dating Quality	Site Type	Source
Camp 11 Lake	MI	46 40 N	88 1 W	549	11	1	Lake	Brubaker, L.B., 1975
Charles Lake	ON	44 44 N	81 1 W	219	1	4	Lake	Bailey, R.E., unpubl.
Chatsworth Bog	IL	40 40 N	88 20 W	219	8	1	Bog	King, J.E., 1981
Chippewa Bog	MI	43 7 N	83 15 W	262	4	2	Bog	Bailey, R.E. & P.J.Ahearn, 1981
Chism-I	PQ	54 48 N	76 9 W	340	1	3	Lake	Richard, P., 1979
Chism-II	PQ	53 5 N	76 19 W	273	1	4	Lake	Richard, P., 1979
Clear Lake	IN	41 39 N	86 32 W	244	5	5	Lake	Bailey, R.E., 1972
Cookstown Bog	ON	44 13 N	79 37 W	234	1	5	Bog	Karrow, P. et al., 1975
Cranberry Glades	WV	38 12 N	80 17 W	1029	3	1	Bog	Watts, W.A., 1979
Crawford Bog	ON	43 28 N	79 57 W	279	0	7	Bog	McAndrews, J.H., unpubl.
Crider's Pond	PA	39 58 N	77 33 W	290	4	5	Lake	Watts, W.A., 1979
Crystal Lake	MI	43 15 N	84 55 W	260	2	7	Lake	Kapp, R.O. and Lay, D. unpubl.
Crystal Lake	PA	41 33 N	80 22 W	313	1	4	Lake	Walker, P.C. & R.I.Hartman, 1960
Cycloid Lake	SK	55 16 N	105 16 W	369	2	2	Lake	Mott, R.J., 1973
Daumont Lake	PQ	54 52 N	69 24 W	600	4	3	Lake	Richard, P., et. al., 1982
Deer Lake Bog	NH	44 2 N	71 50 W	1325	6	2	Bog	Spear, R., 1981
Demont Lake	MI	43 29 N	85 0 W	248	5	2	Lake	Kapp, R.O., 1977
Devils Lake	ND	48 5 N	99 55 W	448	1	4	Lake	McAndrews, J.H., unpubl.
Diana Island	PQ	60 59 N	69 57 W	110	3	2	Lake	Richard, P., 1977
Dismal Swamp (core no. 1)	VA	36 23 N	76 30 W	6	2	3	Bog	Whitehead, D.R., 1972
Disterhaft Farm Bog	WI	43 55 N	89 10 W	329	6	2	Bog	Baker,R.G.1970(Webb & Bryson 1972)
Dosquet	PQ	46 27 N	71 30 W	140	1	4	Bog	Richard, P., 1977
Dufresne	PQ	45 51 N	70 21 W	650	6	2	Lake	Mott, R.J., 1977
Eagle Lake	LB	53 14 N	58 33 W	400	5	1	Lake	Lamb, H.F., 1980
Eagle Lake Bog	NH	44 10 N	71 40 W	1275	3	2	Bog	Spear, R., 1981

TABLE 2: *SITES WITH POLLEN DATA FOR 6000 YR B.P. IN EASTERN NORTH AMERICA.* (Cont'd)

Site Name	State/ Province	Latitude (degrees) (minutes)	Longitude (degrees) (minutes)	Elevation (m)	Number of C-14 Dates	Dating Quality	Site Type	Source
East Baltic Bog	PE	46 26 N	62 7 W	45	3	1	Bog	Anderson, T.W., 1980
Edward Lake	ON	44 22 N	80 15 W	518	3	1	Lake	McAndrews, J.H., 1981
Farnham Bog	PQ	45 17 N	72 59 W	55	4	1	Bog	Morassa,N.,H.Jette,P.Richard,unpub.
Found Lake	ON	45 48 N	78 38 W	488	7	1	Lake	McAndrews, J.H., 1981
Frains Lake	MI	42 20 N	83 38 W	271	9	1	Lake	Kerfoot, W.C., 1974
Gabriel	PQ	46 16 N	73 28 W	250	5	2	Lake	Richard, P., 1977
Gea-I-- St. Hippolyte	PQ	45 59 N	73 59 W	365	1	5	Lake	Richard, P., unpubl.
Glenboro Lake	MB	49 26 N	99 17 W	450	5	6	Lake	Ritchie,J.C. & Lichti-Federovich,S
Goshen Springs	AL	31 43 N	86 8 W	105	4	2	Lake	Delcourt, P.A., 1980
Grand Rapids	MB	53 0 N	98 15 W	350	3	1	Lake	Ritchie, J.C. & K.A.Hadden,1975
Green Lake	MI	44 53 N	85 7 W	305	3	5	Lake	Lawrenz, R., 1975
Grieff Kettle Bog	ON	43 25 N	80 11 W	268	1	7	Bog	Terasmae, J., 1968
Hack Pond	VA	37 59 N	79 0 W	451	2	4	Lake	Craig, A.J., 1969
Harrowsmith	ON	44 25 N	76 42 W	145	1	5	Bog	Terasmae, J., 1968
Hawley Bog Pond	MA	42 34 N	72 53 W	549	5	1	Lake	Patterson, W.A. III, unpubl.
Hayes Lake	ON	49 35 N	93 45 W	391	3	1	Lake	McAndrews, J.H., 1982
Heart Lake	NY	44 11 N	73 58 W	664	8	1	Lake	Whitehead, D.R., unpubl.
Helmetta Bog	NJ	40 23 N	74 26 W	15	1	4	Bog	Watts, W.A., 1979
Horseshoe Lake	MN	45 27 N	93 3 W	331	12	1	Lake	Cushing, E.J., unpubl.
Houghton Bog	NY	42 32 N	78 40 W	428	1	5	Lake	Miller, N.G., 1973
Hudson Lake	IN	41 40 N	86 32 W	239	6	4	Lake	Bailey, R.E., 1972
Iglutalik Lake	BF	66 8 N	66 5 W	90	5	1	Lake	Davis, P.T., 1980
Jacobson Lake	MN	46 25 N	92 43 W	324	4	.	Lake	Wright, H.E., Jr. & W.A.Watts,1969
Jock Lake	ON	48 41 N	86 27 W	290	9	1	Lake	Saarnisto, M., 1975
Joncas Bog	PQ	47 16 N	71 10 W	747	2	1	Bog	Richard, P., 1977
Joncas Lake	PQ	47 15 N	71 10 W	780	1	4	Lake	Richard, P., unpubl.

TABLE 2: *SITES WITH POLLEN DATA FOR 6000 YR B.P. IN EASTERN NORTH AMERICA.* (Cont'd)

Site Name	State/ Province	Latitude (degrees) (minutes)	Longitude (degrees) (minutes)	Elevation (m)	Number of C-14 Dates	Dating Quality	Site Type	Source
Kanaaupscow	PQ	54 1 N	76 38 W	200	1	4	Lake	Richard, P., 1979
Kenogami	PQ	48 22 N	71 34 W	166	2	1	Bog	Richard, P., 1977
Kincardine Bog	ON	44 9 N	81 39 W	198	4	4	Bog	Karrow, P. et al.,1975
Kinsman Pond	NH	44 8 N	71 44 W	1140	3	4	Lake	Spear, R., 1981
Kirchner Marsh	MN	44 50 N	93 7 W	254	6	1	Mars	Wright, H.E., Jr. et al., 1963
Kogaluk Plateau Lake	LB	56 4 N	63 45 W	530	5	2	Lake	Short, S.K. & H.Nichols,1977
Kotiranta	MN	46 43 N	92 37 W	386	3	5	Lake	Wright, H.E., Jr. & W.A.Watts, 196
Lac des Atocas	PQ	45 32 N	73 18 W	120	4	1	Lake	Gauthier, R., 1981
Lac Bouleaux	PQ	45 33 N	73 19 W	126	2	6	Lake	Mott, R.J., unpubl.
Lac Castor	PQ	46 36 N	72 59 W	220	6	1	Lake	Richard, P., unpubl.
Lac Colin	PQ	46 43 N	70 18 W	658	8	1	Lake	Mott, R.J., 1977
Lac Delorme II	PQ	54 25 N	69 55 W	538	4	1	Lake	Richard, P. et.al., 1982
Lac Hamard	PQ	54 48 N	67 30 W	564	4	1	Lake	Stravers, L.K.S., 1981
Lac Louis	PQ	47 17 N	79 7 W	300	3	1	Lake	Vincent, J.S., 1973
Lac Mertini	PQ	47 28 N	72 45 W	242	4	2	Lake	Richard, P., unpubl.
Lac Martyne	PQ	56 47 N	64 50 W	365	3	1	Lake	McAndrews, J.H., unpubl.
Lac Mimi	PQ	47 30 N	70 22 W	411	6	4	Lake	Richard, P., 1977
Lac Romer	PQ	45 58 N	73 20 W	20	4	5	Lake	Comtois, P., 1982
Lac Tanie	PQ	45 46 N	74 18 W	305	6	2	Lake	Richard, P., unpubl.
Lac a la Tortue	PQ	45 32 N	73 19 W	137	4	2	Lake	Gauthier, R., 1981
Lac Yelle	PQ	48 30 N	79 38 W	355	3	2	Lake	Richard, P., unpubl.
Lake Annie	FL	27 12 N	81 21 W	37	8	4	Lake	Watts, W.A., 1975a
Lake of the Clouds	MN	48 9 N	91 7 W	453	30	1	Lake	Craig, A.J., 1972
Lake of the Clouds	NH	44 16 N	71 19 W	1542	5	4	Lake	Spear, R., 1981
Lake E	MB	50 43 N	99 39 W	724	8	1	Lake	Ritchie, J.C., 1969
Lake Louise	GA	30 43 N	83 15 W	61	3	4	Lake	Watts, W.A., 1971

TABLE 2: *SITES WITH POLLEN DATA FOR 6000 YR B.P. IN EASTERN NORTH AMERICA.* (Cont'd)

Site Name	State/ Province	Latitude (degrees) (minutes)	Longitude (degrees) (minutes)	Elevation (m)	Number of C-14 Dates	Dating Quality	Site Type	Source
Lake Mary	WI	46 15 N	89 54 W	488	3	3	Lake	Webb, T. III, 1974
Lake Rogerine	NJ	41 30 N	74 20 W	137	2	4	Lake	Nicholas, J., 1968
Lake West Okoboji	IA	43 22 N	95 11 W	415	10	1	Lake	Van Zant, K.L., 1979
Lanoraie, St. Henri Bog	PQ	45 59 N	73 18 W	18	6	1	Bog	Comtois, P., 1982
,Lantern Hill Pond	CT	41 27 N	71 57 W	36	5	1	Lake	Trent, K.M., unpubl.
"LD" Lake	PQ	50 8 N	67 7 W	122	2	2	Lake	Mott, R.J., 1976
Lima Bog	WI	42 48 N	88 5 W	238	8	2	Bog	Van Zant, K.L., unpubl.
Little Bass Lake	MN	47 17 N	93 36 W	391	5	1	Lake	Swain, P.C., 1979
Little Lake	NB	45 9 N	66 43 W	64	4	5	Lake	Mott, R.J., 1975
Lost Lake	MI	46 43 N	87 58 W	500	9	1	Lake	Brubaker, L.B., 1975
Lynn Lake	MB	56 50 N	101 3 W	340	6	1	Bog	Nichols, H., 1967
Maclaughlin Pond	PE	46 23 N	62 47 W	24	5	2	Lake	Anderson, T.W., unpubl.
Malbaie	PQ	47 36 N	70 58 W	800	2	3	Bog	Richard, P., 1977
Maplehurst	ON	43 13 N	80 39 W	300	6	2	Lake	Mott, R.J. & L.D.Farley-Gill,1978
Marcotte	PQ	47 4 N	71 25 W	503	5	1	Lake	Labelle, C. & P.Richard, 1981
Martin Pond	MN	47 11 N	94 56 W	429	0	7	Lake	McAndrews, J.H., 1966
Mauricie	PQ	46 47 N	72 50 W	270	5	2	Lake	Richard, P., 1977
Mer Bleue	ON	45 24 N	75 30 W	69	1	4	Bog	Mott, R.J. & M.Camfield, 1969
Mingo Pond	TN	35 9 N	86 12 W	300	0	7	Lake	Delcourt, H.R., 1979
Monhegan Island Meadow	ME	43 46 N	69 18 W	3	17	1	Bog	Bostwick, L.K., 1978
Mont Shefford	PQ	45 21 N	72 35 W	282	10	1	Bog	Richard, P., 1977
Mont Valin	PQ	48 36 N	70 50 W	891	4	1	Lake	Richard, P., unpubl.
Montagnais	PQ	47 54 N	71 10 W	800	4	4	Bog	Richard, P., 1977
Moulton Pond	ME	44 37 N	68 38 W	143	16	1	Lake	Davis, R.B. et al, 1975
Muscotah Marsh	KS	39 32 N	95 31 W	320	4	2	Mars	Gruger, J., 1973
Myrtle Lake	MN	47 59 N	93 23 W	393	5	1	Lake	Janssen, C.R., 1968

TABLE 2: *SITES WITH POLLEN DATA FOR 6000 YR B.P. IN EASTERN NORTH AMERICA.* (Cont'd)

Site Name	State/ Province	Latitude (degrees) (minutes)	Longitude (degrees) (minutes)	Elevation (m)	Number of C-14 Dates	Dating Quality	Site Type	Source
Nain Pond	LB	56 32 N	61 49 W	80	8	2	Lake	Short, S.K. & H.Nichols,1977
Napaktok Lake	LB	57 55 N	62 34 W	143	5	2	Lake	Short, S.K., unpubl.
Nelson Pond	MN	46 24 N	92 41 W	335	2	1	Lake	Jacobson, G.L., 1979
North Bay	ON	46 27 N	79 28 W	369	1	4	Bog	Terasmae, J., 1968
North Pond	MA	42 39 N	73 3 W	586	13	1	Lake	Whitehead, D.R. & T.Crisman, 1978
Nunkets Pond	MA	41 58 N	71 3 W	18	4	2	Lake	Bradshaw, R.H.W., unpubl.
Old Field	MO	37 7 N	89 50 W	98	4	1	Bog	King, J.E. & W.H.Allen,Jr.,1979
Palsa Lake	PQ	58 28 N	65 10 W	143	6	1	Lake	Short, S.K., unpubl.
Pamet Cranberry Bog	MA	42 0 N	70 2 W	3	3	2	Bog	Patterson, W.A. III, unpubl.
Panther Run Pond	PA	40 48 N	77 25 W	634	2	4	Lake	Watts, W.A., 1979
Paradise Lake	LB	53 3 N	57 45 W	180	3	2	Lake	Lamb, H.F., 1980
Patricia Bay Lake	BF	70 28 N	68 30 W	11	1	1	Lake	Mode, W., 1980
Perch Lake	ON	46 2 N	77 22 W	160	1	4	Lake	McAndrews, J.H., unpubl.
Pickerel Lake	SD	45 30 N	97 20 W	395	3	2	Lake	Watts, W.A. & R.C.Bright, 1968
Pink Lake	PQ	45 28 N	75 49 W	162	3	2	Lake	Mott, R.J. & L.D.Farley-Gill,1981
Poland Spring Pond	ME	44 2 N	70 21 W	94	11	1	Lake	Jacobson, G.L., unpubl.
Pond Mills Pond	ON	42 55 N	81 15 W	274	1	4	Lake	McAndrews, J.H., 1981
Portage	MN	47 12 N	94 9 W	396	3	2	Lake	McAndrews, J.H., unpubl.
Portage Bog	PE	46 40 N	64 6 W	8	3	3	Bog	Anderson, T.W., 1980
Potts Mountain Pond	VA	37 36 N	80 8 W	840	2	4	Lake	Watts, W.A., 1979
Pretty Lake	IN	41 35 N	85 15 W	294	16	2	Lake	Williams, A.S., 1974
Princeville	PQ	46 8 N	71 56 W	135	0	7	Lake	Richard, P., 1977
Protection Bog	NY	42 37 N	78 28 W	430	3	2	Bog	Miller, N.G., 1973
Pyramid Hills Lake	LB	57 38 N	65 10 W	381	7	1	Lake	Short, S.K. & H.Nichols,1977
Qivituq Cliffs Peat	BF	68 2 N	65 4 W	20	2	1	Bog	Mode, W., 1980
Quicksand	GA	34 19 N	84 52 W	285	2	6	Lake	Watts, W.A., 1970

TABLE 2: *SITES WITH POLLEN DATA FOR 6000 YR B.P. IN EASTERN NORTH AMERICA.* (Cont'd)

Site Name	State/ Province	Latitude (degrees) (minutes)	Longitude (degrees) (minutes)	Elevation (m)	Number of C-14 Dates	Dating Quality	Site Type	Source
Ramsay Lake	PQ	45 36 N	76 6 W	200	5	2	Lake	Mott, R.J. & L.D.Farley-Gill,1981
Riley Lake	ON	54 19 N	84 33 W	142	2	2	Lake	McAndrews, J.H. et al., 1982
Rockyhock Bay	NC	36 10 N	76 41 W	6	13	4	Bog	Whitehead, D.R.,1981
Rogers Lake	CT	41 22 N	72 7 W	91	53	1	Lake	Davis, M.B., 1969
Rossburg Bog	MN	46 35 N	93 36 W	372	4	2	Bog	Wright, H.E., Jr. & W.A.Watts, 196
Round Lake	IN	41 14 N	86 38 W	216	5	5	Lake	Bailey, R.E., unpubl.
Rutz Lake	MN	44 52 N	93 52 W	314	8	2	Lake	Waddington, J.C.B., 1969
Ryerse Lake	MI	46 7 N	85 10 W	259	6	1	Lake	Futyma, R.P., 1982
Sam	PQ	46 39 N	72 58 W	240	6	2	Lake	Richard, P., 1977
Sav-I -- Ste. Agathe	PQ	46 3 N	74 28 W	454	3	3	Lake	Savoie, L. & P.Richard, 1979
Sav-II -- Lac aux Quenoilles	PQ	46 10 N	74 23 W	403	3	4	Lake	Savoie, L. & P.Richard, 1979
Second Lake	ON	44 50 N	73 59 W	196	1	5	Lake	McAndrews, J.H., unpubl.
Shady Valley Peat	TN	36 31 N	81 56 W	383	1	4	Bog	Barclay, F.H., 1957
Shaw's Bog	NS	45 1 N	64 11 W	30	4	1	Bog	Hadden, K.A., 1976
Shouldice Lake	ON	45 9 N	81 25 W	177	2	4	Lake	McAndrews, J.H., unpubl.
Silver Lake	NS	44 33 N	63 38 W	69	3	1	Lake	Livingstone, D.A., 1968
Silver Lake	OH	40 26 N	83 40 W	341	15	2	Lake	Ogden, J.G. III, 1966
Sinkhole Pond	ME	43 58 N	70 21 W	95	12	1	Lake	Jacobson, G.L., unpubl.
St. Benjamin	PQ	46 17 N	70 36 W	330	1	4	Lake	Richard, P., 1977
St. Calixte	PQ	45 57 N	73 52 W	261	4	2	Lake	Richard, P., unpubl.
St. Francois de Sales	PQ	48 17 N	72 8 W	358	4	2	Lake	Richard, P., unpubl.
St. Germain	PQ	45 56 N	74 22 W	473	5	3	Lake	Savoie, L. & P.Richard, 1979
St. Jean, Ile d'Orleans	PQ	46 56 N	70 56 W	68	3	2	Bog	Richard, P., 1977
St. John's Island Pond	LB	53 57 N	58 55 W	137	2	1	Lake	Jordan, R.H., 1975
St. Raymond	PQ	46 53 N	71 48 W	160	2	2	Bog	Richard, P., 1977
Stewart's Dark Lake	WI	45 18 N	91 27 W	335	6	1	Lake	Peters, A. & T.Webb III, 1979

TABLE 2: *SITES WITH POLLEN DATA FOR 6000 YR B.P. IN EASTERN NORTH AMERICA.* (Cont'd)

Site Name	State/ Province	Latitude (degrees) (minutes)	Longitude (degrees) (minutes)	Elevation (m)	Number of C-14 Dates	Dating Quality	Site Type	Source
Stotzel-Leis Site	OH	40 13 N	84 41 W	320	11	1	Bog	Shane, L.C.K., 1976
Szabo Pond	NJ	40 24 N	74 29 W	29	3	4	Lake	Watts, W.A., 1979
Tannersville Bog	PA	41 2 N	75 16 W	277	5	2	Bog	Watts, W.A., 1979
Terhell Pond	MN	47 12 N	95 47 W	442	1	7	Lake	McAndrews, J.H., 1966
Terrien	PQ	46 35 N	70 37 W	404	1	6	Lake	Mott, R.J., 1977
Thompson	MN	47 12 N	96 5 W	370	0	7	Lake	McAndrews, J.H., 1966
Titicut Swamp	MA	41 57 N	71 2 W	18	3	1	Bog	Nelson, S., unpubl.
Tonawa Lake	ON	44 51 N	77 10 W	305	2	2	Lake	McAndrews, J.H., unpubl.
Torrens Bog	OH	40 21 N	82 28 W	302	13	1	Bog	Ogden, J.G. III & R.J.Hay, 1967
Tunturi Lake	PQ	55 1 N	67 30 W	610	3	1	Lake	Stravers, L.K., 1981
Ublik Lake	LB	57 23 N	62 3 W	122	4	1	Lake	Short, S.K. & H.Nichols,1977
Unknown Lake	ME	45 37 N	70 38 W	489	4	7	Lake	Mott, R.J., 1977
Upper South Branch Pond	ME	46 5 N	68 54 W	300	7	2	Lake	Anderson, S. & R.B.Davis, unpubl.
Upper Wallface Pond	NY	44 9 N	74 3 W	945	6	2	Lake	Whitehead, D.R., unpubl.
Val St. Gilles Bog	PQ	49 1 N	79 5 W	290	2	1	Bog	Terasmae, J. & T.W.Anderson,1970
Van Nostrand Lake	ON	44 0 N	79 23 W	297	2	3	Lake	McAndrews, J.H., 1970
Vestaburg	MI	43 25 N	84 53 W	255	4	2	Lake	Gilliam, J.A. et al., 1967
Victoria Road Bog	ON	44 37 N	78 57 W	274	1	4	Bog	Terasmae, J., 1968
Volo Bog	IL	42 21 N	88 11 W	229	6	1	Bog	King, J.E., 1981
Weber Lake	MN	47 28 N	91 40 W	567	4	4	Lake	Fries, M., 1962
West Blue Mound Creek	WI	43 5 N	89 52 W	335	4	1	Bog	Davis, A.M., 1977
White Pond	SC	34 10 N	80 46 W	90	3	4	Lake	Watts, W.A., 1980a
Whitney's Gulch	LB	51 31 N	57 18 W	98	5	2	Lake	Lamb, H.F., 1980
Willow River Pond	MN	46 20 N	92 47 W	314	2	1	Lake	Jacobson, G.L., 1979
Wintergreen Lake	MI	42 24 N	85 23 W	283	8	2	Lake	Manny, B.A. et al., 1978
Woden Bog	IA	43 14 N	93 55 W	381	6	1	Bog	Durkee, L.H., 1971
Wolverine Lake	MI	46 25 N	85 39 W	259	5	1	Lake	Futyma, R.P., 1982

TABLE 2: *SITES WITH POLLEN DATA FOR 6000 YR B.P. IN EASTERN NORTH AMERICA.* (Cont'd)

Site Name	State/ Province	Latitude (degrees) (minutes)	Longitude (degrees) (minutes)	Elevation (m)	Number of C-14 Dates	Dating Quality	Site Type	Source
Wood Lake	WI	45 20 N	90 5 W	464	6	1	Lake	Heide, K.M., 1981
Yamaska	PQ	45 28 N	72 52 W	265	7	1	Lake	Richard, P., unpubl.
Yellow Dog Lake	MI	46 45 N	87 57 W	445	9	1	Lake	Brubaker, L.B., 1975

1: 1 = Bracketing C-14 dates within 2000 years.

2 = Bracketing dates: one within 2000 years and the other within 4000 years.

3 = Bracketing dates within 4000 years.

4 = Bracketing dates: one within 4000 years and the other within 6000 years (where 0 years B.P. at the coretop counts as a date).

5 = Bracketing dates within 6000 years.

6 = Bracketing dates: one within 6000 years and the other within 8000 years.

7 = Non-bracketing dates or no C-14 dates from the core.

TABLE 3: *REGRESSION EQUATIONS FOR MEAN JULY TEMPERATURE FROM THE THIRTEEN CALIBRATION REGIONS IN EASTERN NORTH AMERICA.*

Region A: 54-71 N; 90-110 W

Pollen sum: Alnus + Betula + Cyperaceae + Forb sum + Gramineae + Picea + Pinus

$$\text{July T } (^{o}\text{C}) = \underset{(1.61)}{12.39} + \underset{(.14)}{0.50}{*}\underline{\text{Pinus}}^{.5} + \underset{(.05)}{0.26}{*}\text{Forb sum} = \underset{(.10)}{0.15}{*}\underline{\text{Picea}}^{.5}$$

$$- \underset{(.13)}{0.89}{*}\underline{\text{Cyperaceae}}^{.5} - \underset{(.08)}{0.37}{*}\underline{\text{Gramineae}} - \underset{(.01)}{0.03}{*}\underline{\text{Alnus}}$$

$R^2 = 0.80$; adj. $R^2 = 0.78$; Se = 0.96^{o}C

n = 114; F = 69.86; Pr = 0.0000

Region B: 53-71 N; 50-80 W

Pollen sum: Abies + Alnus + Betula + Herb sum + Picea + Pinus

$$\text{July T } (^{o}\text{C}) = \underset{(2.27)}{8.17} + \underset{(.19)}{0.54}{*}\underline{\text{Picea}}^{.5} + \underset{(.14)}{0.17}{*}\underline{\text{Betula}}^{.5}$$

$$- \underset{(.01)}{0.04}{*}\text{Herb sum} - \underset{(0.01)}{0.01}{*}\underline{\text{Alnus}}$$

$R^2 = 0.70$; adj. $R^2 = 0.70$; Se = 1.52^{o}C

n = 165; F = 95.48; Pr = 0.0000

Region C: 45-55 N; 95-110 W

Pollen sum: Alnus + Betula + Cyperaceae + Forb sum + Gramineae + Picea + Pinus + Quercus

$$\text{July T } (^{o}\text{C}) = \underset{(.77)}{21.80} + \underset{(.09)}{0.29}{*}\underline{\text{Gramineae}}^{.5} - \underset{(.29)}{1.90}{*}\underline{\text{Picea}}^{.25}$$

$$- \underset{(.12)}{0.27}{*}\underline{\text{Alnus}}^{.5} - \underset{(.01)}{0.02}{*}\text{Forb sum}$$

$R^2 = 0.70$; adj. $R^2 = 0.68$; Se = 0.93^{o}C

n = 103; F = 57.33; Pr = 0.0000

TABLE 3: *REGRESSION EQUATIONS FOR MEAN JULY TEMPERATURE FROM THE THIRTEEN CALIBRATION REGIONS IN EASTERN NORTH AMERICA.* (Cont'd)

Region D: 45-55 N; 85-105 W

Pollen sum: Alnus + Betula + Forb sum + Gramineae
+ Picea + Pinus + Quercus + Tsuga

$$\text{July T } (^{o}\text{C}) = \underset{(0.48)}{19.20} + \underset{(.20)}{1.31}{*}\underline{\text{Quercus}}^{.25} + \underset{(.05)}{0.08}{*}\underline{\text{Pinus}}^{.5} + \underset{(.02)}{0.04}{*}\underline{\text{Tsuga}} - \underset{(.06)}{0.38}{*}\underline{\text{Betula}}^{.5} - \underset{(.06)}{0.28}{*}\underline{\text{Picea}}^{.5}$$

$R^2 = 0.79$; adj. $R^2 = 0.77$; $Se = 0.67^{o}C$

$n = 93$; $F = 64.13$; $Pr = 0.0000$

Region E: 4 -5 N; 85-95 W

Pollen sum: Abies + Acer + Alnus + Betula + Carya + Fagus + Fraxinus + Herb sum
+ Juglans + Juniperus + Picea + Pinus
+ Quercus + Tsuga + Ulmus

$$\text{July T } (^{o}\text{C}) = \underset{(0.84)}{18.68} + \underset{(.22)}{1.01}{*}\underline{\text{Quercus}}^{.25} + \underset{(.22)}{0.70}{*}\text{Herb sum}^{.25} + \underset{(.05)}{0.10}{*}\underline{\text{Fraxinus}} - \underset{(.16)}{0.36}{*}\underline{\text{Picea}}^{.25} - \underset{(.16)}{0.33}{*}\underline{\text{Abies}} - \underset{(.07)}{0.21}{*}\underline{\text{Betula}}^{.5} - \underset{(.04)}{0.10}{*}\underline{\text{Fagus}}$$

$R^2 = 0.{*}0$; adj. $R^2 = 0.78$; $Se = 0.65^{o}C$

$n = 108$; $F = 56.79$; $Pr = 0.0000$

TABLE 3: *REGRESSION EQUATIONS FOR MEAN JULY TEMPERATURE FROM THE THIRTEEN CALIBRATION REGIONS IN EASTERN NORTH AMERICA.* (Cont'd)

Region F: 38-46 N; 75-90 W

Pollen sum: Acer + Alnus + Betula + Carya + Fagus + Fraxinus + Gramineae
+ Juniperus + Picea + Pinus + Quercus + Ulmus

$$\text{July T } (^{o}\text{C}) = \underset{(1.10)}{22.22} + \underset{(.12)}{0.30}*\underline{\text{Carya}}^{.5} + \underset{(.08)}{0.20}*\underline{\text{Quercus}}^{.5}$$

$$+ \underset{(.04)}{.14}*\underline{\text{Fraxinus}} - \underset{(.2)}{0.62}*\underline{\text{Pinus}}^{.25} - \underset{(.12)}{0.40}*\underline{\text{Acer}}^{.5}$$

$$- \underset{(.06)}{0.26}*\underline{\text{Betula}}^{.5}$$

$R^2 = 0.69$; adj. $R^2 = 0.68$; $Se = 0.94^{o}C$

$n = 146$; $F = 52.57$; $Pr = 0.0000$

Region G: 40-55 N; 75-87 W

Pollen sum: Abies + Acer + Alnus + Betula + Carya + Fagus + Fraxinus
+ Gramineae + Picea + Pinus + Quercus + Tsuga + Ulmus

$$\text{July T } (^{o}\text{C}) = \underset{(1.09)}{20.77} + \underset{(.26)}{1.09}*\underline{\text{Quercus}}^{.25} + \underset{(.20)}{0.37}*\underline{\text{Carya}}^{.25} + \underset{(.11)}{0.17}*\underline{\text{Fraxinus}}^{.5}$$

$$- \underset{(.24)}{0.81}*\underline{\text{Pinus}}^{.25} - \underset{(.07)}{0.29}*\underline{\text{Picea}}^{.5} - \underset{(.07)}{0.25}*\underline{\text{Alnus}}^{.5} - \underset{(.06)}{0.16}*\underline{\text{Betula}}^{.50}$$

$R^2 = 0.90$; adj. $R^2 = 0.89$; $Se = 0.75^{o}C$

$n = 124$; $F = 154.9$; $Pr = 0.0000$

Region H: 39-48 N; 65-78 W

Pollen sum: Abies + Acer + Alnus + Betula + Carya + Fagus + Fraxinus
+ Juniperus + Myrica + Picea + Pinus
+ Quercus + Tsuga + Ulmus

$$\text{July T } (^{o}\text{C}) = \underset{(0.61)}{17.76} + \underset{(.24)}{1.73}*\underline{\text{Quercus}}^{.25} + \underset{(.02)}{0.09}*\underline{\text{Juniperus}} + \underset{(.15)}{0.51}*\underline{\text{Tsuga}}^{.25}$$

$$- \underset{(.08)}{0.41}*\underline{\text{Picea}}^{.5} - \underset{(.03)}{0.12}*\underline{\text{Acer}} - \underset{(.02)}{0.04}*\underline{\text{Fagus}}$$

$R^2 = 0.77$; adj. $R^2 = 0.76$; $Se = 0.80^{o}C$

$n = 153$; $F = 82.90$; $Pr = 0.0000$

TABLE 3: *REGRESSION EQUATIONS FOR MEAN JULY TEMPERATURE FROM THE THIRTEEN CALIBRATION REGIONS IN EASTERN NORTH AMERICA.* (Cont'd)

Region I: 44-52 N; 60-75 W

Pollen sum: *Abies* + *Acer* + *Alnus* + *Betula* + *Fagus* + *Fraxinus* + *Juniperus* + *Myrica* + *Picea* + *Pinus* + *Quercus* + *Tsuga*

$$\text{July T } (^{o}\text{C}) = \underset{(0.58)}{15.93} + \underset{(.15)}{0.78*\underline{\text{Tsuga}}^{.25}} + \underset{(.23)}{0.73*\underline{\text{Pinus}}^{.25}} + \underset{(.11)}{0.48*\underline{\text{Quercus}}^{.50}}$$

$$+ \underset{(.02)}{0.08*\underline{\text{Abies}}} + \underset{(.02)}{0.06*\underline{\text{Juniperus}}} - \underset{(.07)}{0.19*\underline{\text{Myrica}}^{.5}} - \underset{(.01)}{.05*\underline{\text{Picea}}}$$

$R^2 = 0.74$; adj. $R^2 = 0.72$; Se $= 0.68^{o}$C

$n = 114$; $F = 43.13$; $Pr = 0.0000$

Region J: 48-54 N; 50-75 W

Pollen sum: *Abies* + *Alnus* + *Betula* + *Cyperaceae* + *Picea* + *Pinus*

$$\text{July T } (^{o}\text{C}) = \underset{(3.68)}{15.28} + \underset{(.62)}{1.64*\underline{\text{Pinus}}^{.25}} = \underset{(0.06)}{0.06*\underline{\text{Abies}}}$$

$$- \underset{(0.32)}{0.34*\underline{\text{Picea}}^{.5}} - \underset{(.03)}{0.05*\underline{\text{Cyperaceae}}} - \underset{(.02)}{0.03*\underline{\text{Betula}}}$$

$R^2 = 0.48$; adj. $R^2 = 0.44$; Se $= 1.34^{o}$C

$n = 61$; $F = 10.45$; $Pr = 0.0000$

Region K: 28-39 N; 75-95 W

Pollen sum: *Betula* + *Carya* + *Cupressaceae* + *Cyperaceae* + *Fraxinus* + Herb sum + *Liquidambar* + *Nyssa* + *Pinus* + *Quercus* + *Salix* + *Ulmus*

$$\text{July T } (^{o}\text{C}) = \underset{(0.25)}{26.92} + \underset{(.08)}{0.36*\underline{\text{Liquidambar}}^{.5}} + \underset{(.01)}{.008*\underline{\text{Cupressaceae}}}$$

$$+ \underset{(.00)}{.001*\underline{\text{Pinus}}} - \underset{(.12)}{0.58\ \underline{\text{Betula}}^{.5}} - \underset{(.00)}{.009*\underline{\text{Quercus}}}$$

$R^2 = 0.49$; adj. $R^2 = 0.47$; Se $= 0.72^{o}$C

$n = 115$; $F = 21.31$; $Pr = 0.0000$

TABLE 3: *REGRESSION EQUATIONS FOR MEAN JULY TEMPERATURE FROM THE THIRTEEN CALIBRATION REGIONS IN EASTERN NORTH AMERICA.* (Cont'd)

Region L: 28-39 N; 80-91.5 W

Pollen sum: *Betula* + *Carya* + *Cupressaceae* + *Cyperaceae* + *Fraxinus* + Herb sum
+ *Liquidambar* + *Nyssa* + *Pinus* + *Quercus*
+ *Salix* + *Ulmus*

$$\text{July T } ({}^{o}\text{C}) = \underset{(0.30)}{26.48} + \underset{(.10)}{0.40}{*}\underline{\text{Liquidambar}}^{.5} + \underset{(.06)}{0.16}{*}\underline{\text{Cupressacea}}^{.5}$$

$$+ \underset{(.00)}{.005}{*}\underline{\text{Pinus}} - \underset{(.15)}{0.62}{*}\underline{\text{Betula}}^{.5} - \underset{(.00)}{.006}{*}\underline{\text{Quercus}}$$

$R^2 = 0.59$; adj. $R^2 = 0.56$; $Se = 0.72^{o}C$

$n = 82$; $F = 21.56$; $Pr = 0.0000$

Region M: 25-35 N; 80-87 W

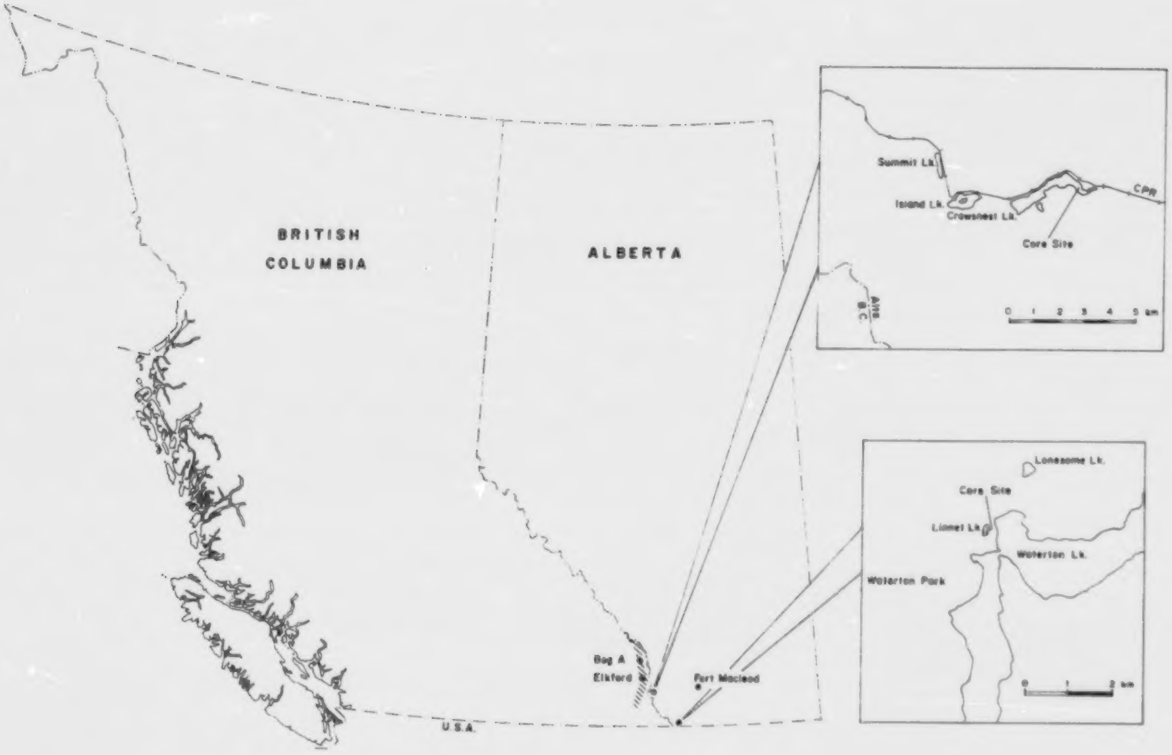
Pollen sum: *Carya* + *Cyperaceae* + *Cupressaceae* + Herb sum
+ *Liquidambar* + *Pinus* + *Quercus*

$$\text{July T } ({}^{o}\text{C}) = \underset{(0.38)}{26.49} + \underset{(.10)}{0.23}{*}\underline{\text{Cupressaceae}}^{.5} + \underset{(.00)}{0.01}{*}\underline{\text{Pinus}}$$

$$- \underset{(.01)}{0.02}{*}\underline{\text{Quercus}}$$

$R^2 = 0.57$; adj. $R^2 = 0.52$; $Se = 0.36^{o}C$

$n = 35$; $F = 11.90$; $Pr = 0.0000$


BRITISH COLUMBIA
ALBERTA
U.S.A.
Bog A
Elkford
Fort Macleod
Summit Lk.
Island Lk.
Crowsnest Lk.
CPR
Core Site
Alta.
B.C.
0 1 2 3 4 5 km
Lonesome Lk.
Core Site
Linnet Lk.
Waterton Lk.
Waterton Park
0 1 2 km

FRONTISPIECE: *General locality map for Parts 1-3 of this paper.*

POSTGLACIAL POLLEN AND PALEOCLIMATE IN SOUTHWESTERN ALBERTA AND SOUTHEASTERN BRITISH COLUMBIA[1]

L.V. Hills[2], O.A. Christensen[3], A. Fergusson[4], J.C. Driver[5], and B.O.K. Reeves[6]

PART 1: PALYNOLOGIC AND PALEOCLIMATIC INTERPRETATION OF HOLOCENE SEDIMENTS, WATERTON LAKES NATIONAL PARK, ALBERTA

O.A. Christensen and L.V. Hills

INTRODUCTION

The sampling of several lake bottoms in Waterton Lakes National Park for palynomorphs is part of a wider program initiated by the National and Historic Parks Branch to carry out a broad resource inventory of this park. Of the several bodies of water examined in the park within the subalpine zone, Linnet Lake (el. 1299 m) was chosen because of the absence of streams entering the lake, and its favourable location adjacent to the forest-grassland transition. Linnet Lake, an ice block depression, is located adjacent to the north end of upper Waterton Lake in Waterton Lakes National Park.

Volcanic tephra (tentatively identified as Mazama tephra (6600 yr B.P.)) is present at the bottom of the Linnet Lake cores, and two radiocarbon dates; 4935 ± 220 yr B.P. (GX 2476) from 1.9 m, and 2830 ± 150 yr B.P. (GX 3538) from 0.35 m provide temporal control on the core. The oldest radiocarbon date closely approximates the initiation of the first cold period following the Altithermal, whereas the youngest one dates the second period of cooling which apparently continues to the present time. Therefore, the core encompasses approximately the last 7000 years.

Waterton Park lies on a storm track resulting in the presence of maritime air masses, causing warmer weather and increased winter precipitation. The prevailing wind direction for most of the year, however, is southerly. The lowest mean temperatures occur from

---

[1] Frontispiece (opposite) is a general locality map for Parts 1 - 3 of this paper.

[2] Department of Geology and Geophysics, University of Calgary, Calgary, Alberta T2N 1N4

[3] Deceased.

[4] PanCanadian Petroleum Ltd., Box 2850, Calgary, Alberta T2P 2S5

[5] Department of Anthropology, Simon Fraser University, Burnaby, British Columbia V5A 1S6

[6] Department of Archaeology, University of Calgary, Calgary, Alberta T2N 1N4

November to January, and maximum precipitation occurs in three periods: August to September, December to January and April to June (Milne-Brumley 1971).

Linnet Lake is a kettle lake separated from Waterton Lake by porous gravels such that the water level in the former is controlled by the water level in the latter. The level of Waterton Lake in turn is related to runoff and to an actively growing fan at the lower end of the lakes (personal communication, J. Harrison, 1971).

The lowest yearly level of Waterton Lakes is probably reached in late winter - early spring, whereas the highest level is reached from late May to early July (Milne-Brumley 1971). The maximum height would vary, reaching a peak during flood years. These peak flood years may be represented in the core by the coarse sediment layers which probably represent silt and sand derived from outwash gravels by wave action.

Linnet Lake is located in the subalpine zone of the coniferous forest. Vegetation is of two general types, Boreal and Cordilleran. Boreal species occur across northern Canada, whereas species with Cordilleran distribution are confined to the mountain region of western North America (Ogilvie, no date). Characteristically, the arboreal element of the subalpine zone in Waterton Park is dominated by white spruce (*Picea glauca* (Moench) Voss) and Engelmann spruce (*Picea engelmannii* Parry). Paper birch (*Betula papyrifera* Marsh) along with some Douglas fir (*Pseudotsuga menziesii* (Poir) Britt) found in drier areas, occur throughout the lower and middle part of the subalpine zone. Aspen (*Populus tremuloides* Michx) is also common in these parts of the subalpine zone, especially where forest approaches the grasslands. Poplar (*Populus trichocarpa* Torri and Gray) and lodgepole pine (*Pinus contorta* var. *latifolia* Engelm) occur at all elevations below timberline. The latter, however, is only fire successional in the Canadian Rockies (Ogilvie, no date). Typical upper timberline species include: alpine fir (*Abies lasiocarpa* (Hook) Nutt), whitebark pine (*Pinus albicaulis* Engelm) and larch (*Larix lyalli* Parl.).

FIELD PROCEDURES

Samples were taken at the southern end of the lake in water reaching depths of 2.2 m. These were recovered with a Livingstone (1955) sampler, a variant of a device originally designed by Dachnowsky (Kapp 1969). Sampling was carried out from a 1.9 x 2.5 m raft, constructed from two by sixes with 61 x 20.3 x 152.4 cm styrofoam pads at each end to increase its buoyancy. Sampling was done through a 30.5 x 15.2 cm hole in the centre of the raft. Anchors were placed at both ends to ensure stability. The walls of the sampling hole were protected from collapse by the insertion of a 50.8 cm length of drain pipe. Samples were extruded in the field into lengths of polyethylene tubing and transported to the laboratory the same day.

LABORATORY PROCEDURES

Samples were prepared by initial treatment with 52% HF to remove inorganic materials. Following this, the acetolysis technique outlined by Erdtman (1960) was used. The remaining organic material was treated by boiling in 10% K_2CO_3 and differential centrifuging. Palynomorphs were permanently embedded in polyvinylalcohol and mounted in Lakeside 70 without staining.

STRATIGRAPHY

In the lower level in both cores taken, boulders prevented further penetration of the sampler. The major part of the cores consisted of a light-to dark-brown gyttja, some levels of which contained larger plant remains. The gyttja bands were separated by bands of grey "silt" or, less commonly, coarse sand. Examination of the "silt" bands suggests that a number of these contained volcanic glass shards, and presumably represent post-Mazama tephra falls or reworked Mazama tephra. The coarse sand interbeds may have formed during periods of high water, which could have washed the sand out of the gravels surrounding the kettle or may be the result of run off during heavy rains or snowmelt.

THE POLLEN DIAGRAM

Palynomorph frequencies in each level were derived from a count of 300 or more grains per level. The palynomorphs of the pollen sum included all terrestrial tree, shrub and herb types. Semi-aquatics and aquatics were excluded from the sum but are recorded in the pollen diagram as percentages of the pollen sum. Relative, rather than absolute frequencies, were used, since changes in sedimentation rate as indicated by changes in sediment type made it impossible to sample according to a predetermined time interval.

Pollen Zones

Four zones are suggested for the Linnet Lake core (Figure 1). The division into four zones is largely based upon the relative frequencies of *Pinus, Picea, Abies*, and the grassland elements (Gramineae and *Artemisia*).

Zone IV (see Figure 1, sample numbers 10-10 to 9-5), the lowest zone, is dominated by *Pinus* (40-80%) associated with high frequencies of *Populus*, Gramineae, Chenopodiineae, *Ambrosia* and *Artemisia*.

Zone III (sample numbers 9-5 to 8-1) differs from Zone IV by a general decrease in *Pinus, Populus*, Gramineae, Chenopodiineae, *Ambrosia* and *Artemisia* over the previous zone. *Abies* increases notably and reaches a maximum frequency in this zone. On the whole, arboreal pollen reaches a maximum frequency relative to shrubs and herbs within this zone (Figure 1).

Zone II (sample numbers 8-1 to 7-4) is dominated by *Pinus*. *Picea* decreases from the previous period. *Betula* and *Alnus* increase over the previous period. Gramineae, Chenopodiineae, *Ambrosia* and *Artemisia* also increase in frequency. Frequencies from Zone II are highly reminiscent of Zone IV.

Zone I (sample numbers 7-4 to 7-1) is marked by an increase in *Picea* pollen which becomes the dominant contributor to the arboreal pollen. *Pinus* and Gramineae decrease and stabilize in frequency. *Betula, Alnus* and *Corylus* increase and then decrease toward the top of the section.

INTERPRETATION

Because of the lack of work on recent pollen spectra in the Canadian Rockies, it is difficult to use the comparative method (Ritchie and Lichti-Federovich 1968, p. 877) in an interpretation of the Linnet Lake core. By the comparative method, we mean the relating of pollen assemblages within various levels of the core to recent pollen samples that characterize modern vegetational types. These vegetational types would in turn reflect various kinds of landforms and climatic phenomena. Thus, most palaeoecological and vegetational postulates are of a tentative nature. The work of Baker (1970) and Waddington and Wright (1970) in Yellowstone Park, Montana, provides a basis for interpretation.

The period of time immediately following the Pleistocene is not represented in the Linnet Lake core. Both cores taken in the lake were characterized by a unit of volcanic

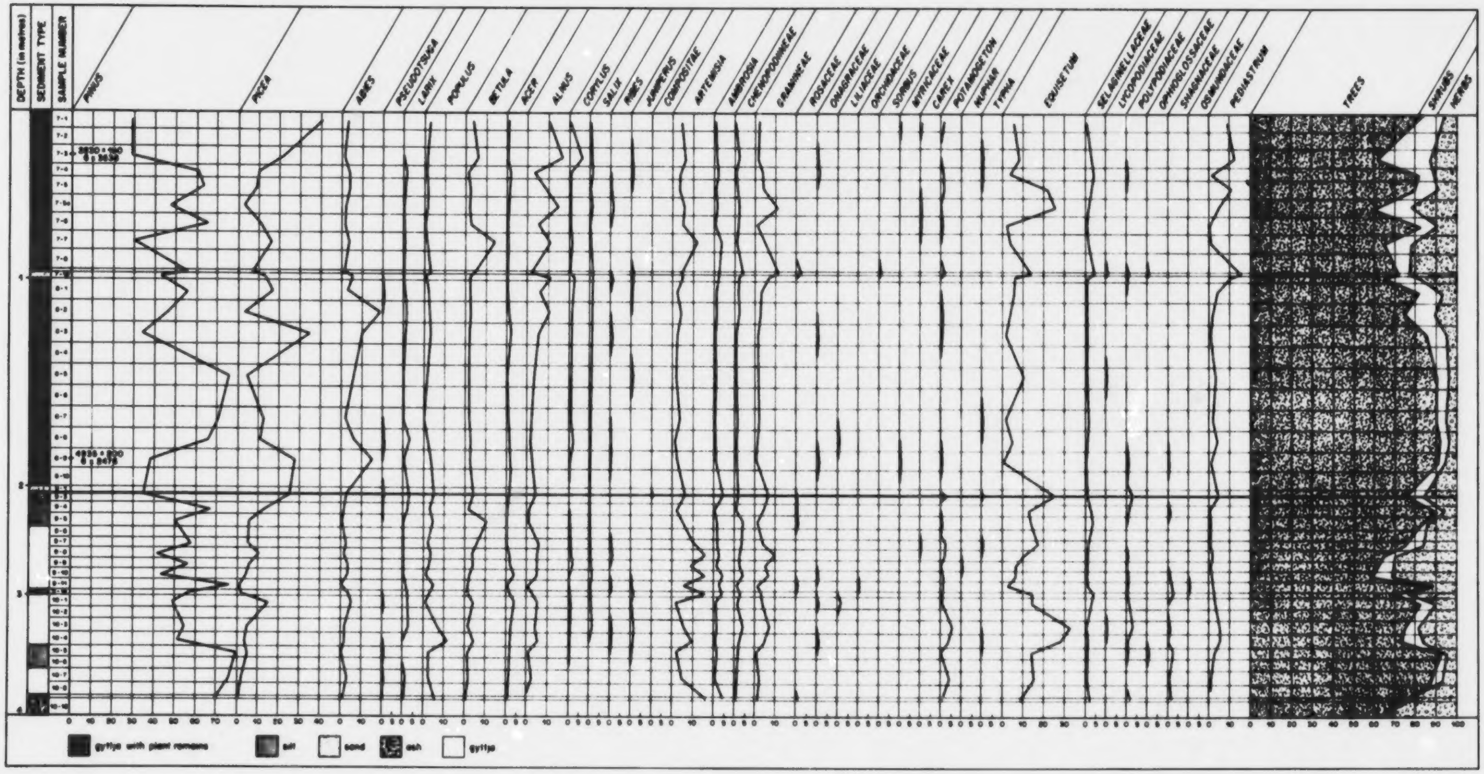


FIGURE 1: *Pollen diagram for Linnet Lake, Waterton Lakes National Park, Alberta.*

tephra or volcanic tephra and interbedded silt at the bottom. This is probably Mazama tephra, dated to 6600 yr B.P. and thus provides time control for the base of the core.

Zone IV

Changes throughout Zone IV are probably largely related to successional changes in *Pinus contorta*, which is fire successional in this area. The assemblage as a whole is indicative of a drier and/or warmer period than at present. The larger amounts of *Betula, Acer* and *Alnus* suggest that the period may be warmer rather than drier, unless a lowering of the level of Linnet Lake is postulated, leading to a wider expanse of hygrophytic woodland around the exposed lake margin. However, the lake margin is steep, and therefore, lowering unless almost complete would provide only a small additional area for this vegetation type.

This zone can probably be related to the level recorded by Hansen (1948) at Fish and Johns Lakes (Glacier Park, Montana) near a volcanic ash band in which Yellow pine (*Pinus pondersoa* Laws.) and Douglas fir (*Pseudotsuga menziesii*) attain their maximum. Since Yellow pine and Douglas fir both favour xeric sites, this, along with a grass-chenopod-composite maximum is taken to suggest a period of drying.

In Jasper National Park, Heusser (1956) notes a warmer, drier period at what is probably the Mazama tephra level, marked by a decline of *Abies*, and a *Pseudotsuga menziesii* peak. East of Jasper, Hansen (1949) notes a *Pinus* maximum at the ash level at Edson and Entwhistle bogs. This he suggests indicates a decrease in moisture. Perhaps this zone can also be correlated with Lichti-Federovich's (1970) Zone L-4 in central Alberta, characterized by open mixed wood with grassland on xeric sites and rare boreal elements. The termination of Zone IV is in agreement with the findings of MacDonald (1982) and Mott and Jackson (1982). Zones III through I correspond to an undifferentiated interval within the latter two studies. Probably this reflects their location well within a floral zone as compared to the boundary position of Linnet Lake.

A warm, dry interval is also suggested at this level by Ritchie and Lichti-Federovich (1968) in Manitoba.

Zone III

The major change from Zone IV to Zone III (ca. 5,000 yr B.P.) is an increase in *Abies*, a more cold-tolerant species than *Pinus contorta* or *Picea glauca*. Although the *Abies* increase appears as two peaks, likely the separation of these reflects a relative increase in *Pinus*, due to a sustained period of forest fire, rather than a climatic change. This would have the effect of appearing to lower the relative amount of *Abies* pollen, whereas the absolute amount may not decrease greatly. This zone probably represents a deterioration of the climate of Zone IV (cooler and moister). The decrease in *Alnus, Acer, Populus* and grassland species also seems to suggest a cooling trend. A radiocarbon date of 4935 ± 220 yr B.P. (GX 2476) for the initiation of this phase indicates that, temporally, it agrees with the end of the Altithermal in Yellowstone Park (Baker 1970; Waddington and Wright 1970). The decrease in Gramineae and increase in *Pinus* and *Picea* at 5200 yr B.P. (Lichti-Federovich 1970) probably reflects this same change in climate.

In Fish and Johns Lakes, Glacier Park, Montana, this climatic deterioration is reflected in the increase of White Pine (*Pinus monticola* Dougl.), *Picea* and *Abies* over the previous period (Hansen 1948). In Jasper, following the Douglas fir peak, Heusser (1956) records an increase in *Abies, Picea* and *Tsuga*, reflecting a cooler climate. At Edson and Entwhistle Bogs, *Picea* and *Abies* increase above the volcanic ash level.

In Manitoba, Ritchie (1969), using absolute counts, suggested a climatic deterioration about 2500 years ago; whereas relative counts (Ritchie and Lichti-Federovich 1968) suggest that this deterioration started earlier, just before 3570 years ago. McAndrews (1966) suggests a similar event for northwestern Minnesota, marked by the replacement of Oak Savanna by Mesic Deciduous Forest (*Quercus-Ostrya* assemblage zone) at about 4000 years ago.

Zone II

Picea and *Abies* decrease in Zone II (ca. 3000 to 4000 years ago), whereas *Pinus* is still the major contributor to the arboreal pollen. An increase in *Betula, Alnus* and the grassland elements suggest a warming trend, or, in the case of the latter, an increase in xeric sites or more open woods. This zone may be related to increases in *Pinus* and decreases in *Picea* at Edson and Entwhistle Bogs (Hansen 1949) and to the slight increase in Gramineae at that locality.

Zone I

Zone I (ca. 3000 years ago to present) is characterized by a progressive increase in, and establishment of, present-day proportions of *Picea*. This increase in spruce is basically at the expense of *Pinus*. *Betula, Alnus* and *Corylus* initially increase in this period and then decrease, suggesting a possible increase in moisture, followed by cooling or perhaps simply a successional change due to a period with few fires.

DISCUSSION

The Linnet Lake core reflects an initial period of warm, dry conditions (Altithermal) with Lodgepole pine forest covering most of the area. Drier sites probably contained grassland elements and perhaps aspen, whereas more mesic sites contained *Betula, Acer* and *Alnus*. This was followed about 5000 years ago by a period of deteriorating climate, allowing the *Picea* and *Abies* component to increase. Cooling of the climate may have resulted also in the lowering of the altitudinal range for *Abies*, accounting for the peak at this level. The forest was apparently closed then, since the arboreal pollen reaches its maximum compared to shrubs and herbs.

This interval was followed by a period of warmer, drier climate, which resulted in considerable expansion of grassland (0.5 to 1.0 m depth) into the area - possibly more extensive than in the interval 7000 to 5000 yr B.P. This phase may have begun about 4000 yr B.P. and terminated about 3000 yr B.P. However, the former date cannot be corroborated at the present time. A cooling trend, starting about 3000 years ago, led to the vegetation achieving present-day proportions - a spruce-dominated forest with some *Betula, Alnus* and *Corylus* in moister sites.

Throughout the pollen record (Figure 1), a gradual change consisting of decreasing *Pinus* and grassland elements and increasing *Picea*, can be recognized. The decreases in *Pinus* are probably related to normal succession from *Pinus* to *Picea* and *Abies*-dominated stands. The sharp increase of *Pinus* at various levels reflects periods of fire. These fires were probably not widespread since the Onagraceae pollen (fireweed) never reaches high proportions in any level. Thus, the core from Linnet Lake records: a warm dry interval (Altithermal) from 7000 to 5000 yr B.P.; a cooler period with some downward movement of such alpine species as *Abies* between 5000 and ?4000 yr B.P.; a warm, dry period with considerable expansion of the grassland from ?4000 to 3000 yr B.P.; and a cooling trend from 3000 yr B.P. to the present time.

Osborn (1982) summarized evidence for three Holocene glacial events within the southern Canadian Rocky Mountains. The oldest (Crowfoot Advance) predates the Mazama tephra and is therefore older than the base of the Linnet Lake core. An intermediate event, characterized by rock glacier activity at Lost Horse Creek in Banff National Park, may correlate with the climatic cooling reflected in the Linnet Lake core between approximately 5000 and 4000 yr B.P. The youngest (Cavell Advance) dates principally to the last 400 years and represents the culmination of a climatic trend started about 3000 years ago.

SUMMARY

A sequence of palynological events spanning the last 7000 years is developed for the Waterton Lakes area. The period from 7000 to 5000 yr B.P. was warmer and drier than present, from 5000 to ?4000 yr B.P. a cooler, moister climate prevailed, followed by a warmer and drier climate from ?4000 to 3000 yr B.P., which in turn was followed by a progressive cooling of the climate from 3000 yr B.P. to the present.

ACKNOWLEDGEMENTS

Financial support of the National and Historic Parks Branch, Canada Department of Indian Affairs and Northern Development, the National Museum of Natural Sciences Climatic Change Project and a Natural Sciences and Engineering Research Council of Canada grant to L.V. Hills made this research possible. The help of Mrs. B. McCaffery, Mrs. M. Hills and Tari Forrest in sample preparation and typing is gratefully acknowledged. This paper is based on a report submitted to the National and Historic Parks Branch, Canada Department of Indian Affairs and Northern Development by O.A. Christensen and L.V. Hills (1971).

REFERENCES

Baker, R.G. 1970. Late-Quaternary vegetation history in Yellowstone Park. First Meeting, American Quaternary Association, Yellowstone Park and Montana State University, Bozeman. Abstracts: 4.

Christensen, O.A., and L.V. Hills. 1971. Palynologic and paleoclimatic interpretation of Recent sediments, Waterton Lakes National Park, Alberta. Report to National and Historic Parks Branch, Department of Indian Affairs and Northern Development. 15 pp.

Erdtman, G. 1960. The acetolysis method. Svensk Botaniska Tidskrift 54:561-564.

Hansen, H.P. 1948. Postglacial forests of the Glacier National Park region. Ecology 29:146-152.

______. 1949. Postglacial forest in west central Alberta, Canada. Bulletin of the Torrey Botanical Club 76:278-289.

Heusser, C.J. 1956. Postglacial environments in the Canadian Rocky Mountains. Ecological Monographs 26:263-302.

Kapp, R.O. 1969. Pollen and spores. Wm. C. Brown Company, Dubuque, Iowa. 249 pp.

Lichti-Federovich, S. 1970. The pollen stratigraphy of a dated section of Late Pleistocene lake sediment from central Alberta. Canadian Journal of Earth Sciences 7:938-945.

Lichti-Federovich, S., and J.C. Ritchie. 1965. Contemporary pollen spectra in Central Canada II, The forest grassland transition in Manitoba. Pollen et Spores 7:63-87.

______. 1968. Recent pollen assemblages from the western interior of Canada. Review of Paleobotany and Palynology 7:297-344.

MacDonald, G.M. 1982. Late Quaternary paleoenvironments of the Morley Flats and Kananaskis Valley of southwestern Alberta. Canadian Journal of Earth Sciences 19(1):23-25.

McAndrews, J.H. 1966. Postglacial history of prairie, savanna and forest in northwestern Minnesota. Torrey Botanical Club Memoirs 22(2):1-72.

Milne-Brumley, L. 1971. The Narrows Site in Waterton Lakes National Park, Alberta. M.A. thesis, University of Calgary, Calgary, Alberta.

Mott, R.J., and L.E. Jackson, Jr. 1982. An 18,000-year palynological record from the southern Alberta segment of the classical Wisconsin "Ice-Free" Corridor. Canadian Journal of Earth Sciences 19:504-513.

Ogilvie, R.T. (no date). Ecology of vegetation in Banff National Park. (Manuscript filed in Department of Biology, University of Calgary).

Osborn, G. 1982. Holocene glacier and climate fluctuations in the southern Canadian Rocky Mountains: a review. Striae 18:15-25.

Ritchie, J.C. 1969. Absolute pollen frequencies and carbon-14 age of a section of Holocene lake sediment from the Riding Mountain area of Manitoba. Canadian Journal of Botany 47:1345-1349.

Ritchie, J.C., and S. Lichti-Federovich. 1968. Holocene pollen assemblages from the Tiger Hills, Manitoba. Canadian Journal of Earth Sciences 5:873-880.

Waddington, J.C.B., and H.E. Wright, Jr. 1970. Late-Quaternary vegetational changes on the east side of Yellowstone Park, Wyoming. First Meeting, American Quaternary Association, Yellowstone Park and Montana State University, Bozeman. Abstracts: 139.

PART 2: A PALYNOLOGICAL RECORD, UPPER ELK VALLEY, BRITISH COLUMBIA

Angus Fergusson and L.V. Hills

INTRODUCTION

Quaternary palynological studies of southwestern Alberta have recently been enhanced through publications by MacDonald (1982) and Mott and Jackson (1982). Prior to this the work of Alley (1976), Hansen (1949A, 1949B, 1955), Harrison (1976), Heusser (1956, 1960) and Lichti-Federovich (1970, 1972) served as the published base for the area of southwestern Alberta and southeastern British Columbia. In addition unpublished data of Alley (1972), Christensen and Hills (1971, this volume), Driver (1978), Fergusson (1978) and Bujak (1974) have been used. The purpose of this paper is to report on a bog profile from the upper Elk Valley (Fergusson 1978).

STUDY AREA

The peat bog considered in this study (informally referred to as Bog A) is located 41.5 km north of Elkford, British Columbia (Figure 1; Latitude 50° 23.5'N, Longitude 114° 56.0'W; see also Frontispiece at the beginning of this series of papers) and at an elevation of 1586m. The bog contains approximately 5 m of sediment and is located in a depression on hummocky moraine. The moraine is composed of a clayey silt till that was deposited along the valley side during glacial retreat (Fergusson 1978).

Local vegetation at Bog A consists of spruce (*Picea engelmanni* Parry) with a shrub cover of willow (*Salix*), Labrador tea (*Ledum groenlandicum* Oeder), *Betula glandulosa* (Michx), and a herb cover of sedges (Cyperaceae), valerian (*Valeriana*) and fleabane (*Erigeron*) (personal communication, D. Polster, 1977).

Three sediment cores, each 4.2 m in length, were collected from the bog using a Swedish Hiller peat sampler (Faegri and Iversen 1965) in July, 1977. Sediment samples taken at 10-cm intervals, each approximately 1 cm long, were collected from one of the cores. The remaining cores were used to determine sediment stratigraphy. Debris left by an excavation into the bog was examined and compared to the sediment stratigraphy determined by the core samples. Samples of tephra, gyttja, and inorganic clay were collected from the pit for examination in the laboratory.

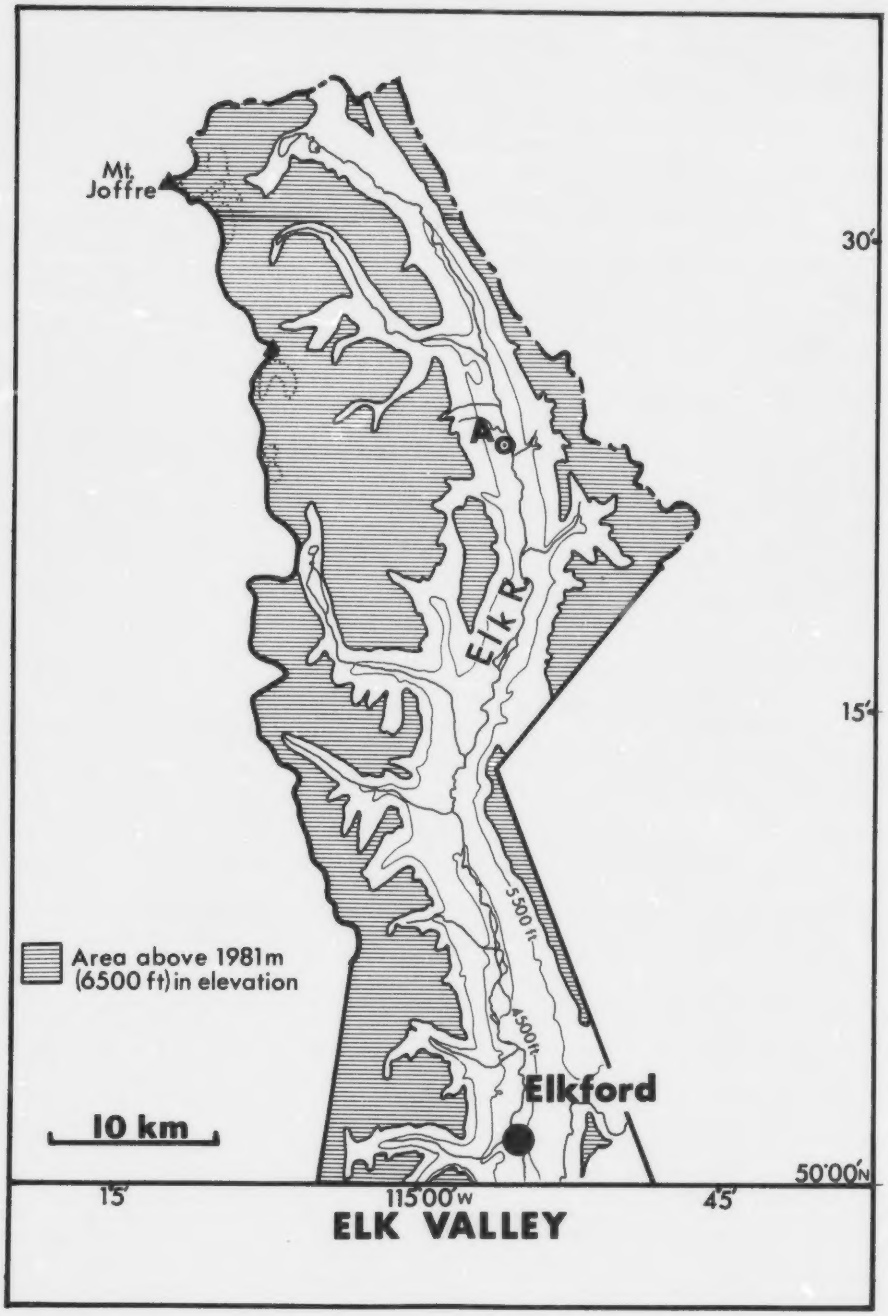

Mt. Joffre
A
Elk R.
5500 ft
4500 ft
Elkford
Area above 1981m (6500 ft) in elevation
10 km
30'
15'
50°00'N
15'
115°00'W
45'
ELK VALLEY

FIGURE 1: *Location of Bog A, upper Elk Valley, British Columbia.*

Forty-two samples from the upper 3.3 m of the core in the peat layer, were washed through a 60 mesh sieve to separate out the coarser organic particles. The fines were washed into a 400 ml beaker. After allowing 60 seconds for the particles to settle, the remaining suspended clays were decanted. This was repeated after stirring until the water remained clear. In order to remove all the carbonates, 10% HCl was added until effervescence ceased. All samples were then treated overnight with HF to remove silicates, then oxidized for 10 minutes with Schulze solution (KNO_3 + $KClO_3$). The oxidized humic compounds were then dispersed with K_2CO_3 for 20 minutes (Gray 1965). The samples were thoroughly washed between each step. Safranin O was used to stain the grains before mounting. The pollen concentrations were mixed in polyvinyl alcohol on cover slips and allowed to dry. They were then mounted on slides with Lakeside 70 cement.

The relative frequency, for each sample, was determined after identifying and counting 250 grains. A percentage pollen diagram was then constructed to show variations in the frequencies with time. The upper 10 cm of sediment was not examined due to lack of sediment recovery. There were only a few identifiable pollen grains in the clay interval from 4.2 to 3.6 m, therefore no pollen frequencies could be determined.

The main pollen identifications were made by referring to Kapp (1969) and Weir and Thurston (1976), and finally by direct comparison with modern pollen. The distinction between larch (*Larix*) and Douglas fir (*Pseudotsuga*) was difficult because the pollen grains are similar. They were separated mainly on size.

RESULTS

Sediment Stratigraphy

Three stratigraphic units occur in the core: (1) a lower clay from 4.2 to 3.6 m; (2) a gyttja layer from 3.6 to 3.3 m; and (3) a peat layer from 3.3 to 0 m. The peat layer is separated at 1.59 to 1.53 m by Mazama tephra.

The clay layer is dark gray (10 YR 4/1) when moist. The lower 0.2 m of the unit is gritty and grades into underlying till of the hummocky moraine. Striated pebbles were found in the lower section of the clay. The clay from 4.0 to 3.6 m contains gastropods, pelecypods, ostracodes and beetle remains. The freshwater mollusc shells are well preserved and articulated indicating a quiet water environment of deposition.

The gyttja layer is very dark grayish brown (10 YR 3/2) when moist. The unit contains articulated pelecypods and ostracodes, beetles, and complete leaves.

The peat deposits consist of layers of spongy mosses and layers of more compact wood fragments. A unit of peat at the peat/gyttja contact contained beaver-chewed wood. The tephra, pale yellow (2.5 Y 7/4) when dry, can easily be distinguished from the overlying and underlying peat deposits.

Dating

Fergusson and Osborn (1981) provide three radiocarbon dates for the core. The lowermost, 13,430 ± 450 yr B.P. (GX 5599; C^{13}-corrected) was based on gastropods collected between 4.5 and 3.6 m. The second date of 10,125 ± 285 yr B.P. (GX 5598) was obtained from beaver-chewed wood at 3.3 m depth. The mazama tephra (6600 yr B.P.; Reubin and Alexander 1960) from 1.59-1.53 m, identified by microprobe analysis of titanomagnetites and refractive indices of glass shards (Fergusson 1978) provides the third date.

Pollen Stratigraphy

The sediment core is divided into five pollen zones. Zones III and V each have a subzone that is recognized by a decrease in the pine to spruce plus fir ratio (p/sf) ratio (Figure 2). Zone I, dating between 13,400 and than 10,000 yr B.P., is characterized by a high percentage of shrubs and herbs consisting of birch (*Betula*: 9-20%), sagebrush (*Artemisia*: 7-12%), sedge (Cyperaceae: 0-7%), willow (*Salix*: 1.6%) and minor amounts of alder (*Alnus*), soapberry (*Shepherdia*), chenepod (Chenopodiaceae), ragweed (Ambrosiaceae), heath (Ericaceae) and horsetail (*Equisetum*). The tree pollen consists of spruce (*Picea*: 4-10%) and pine (*Pinus*: 50-60%). Pine is dominant throughout the profile.

Zone II is distinguished by an increase in spruce (19-24%) and pine (56-71%). The p/sf ratio decreases dramatically from Zone 1. Sagebrush (2%) and birch (1-4%) have greatly decreased. Moss (0-5%) and sedge (1-4%) are present at first, then suddenly decrease.

Zone III is dominated by pine (82-89%) and spruce (6-12%). The p/sf ratio is increased. Herb and shrub percentages are notably decreased. Subzone IIIa is similar to

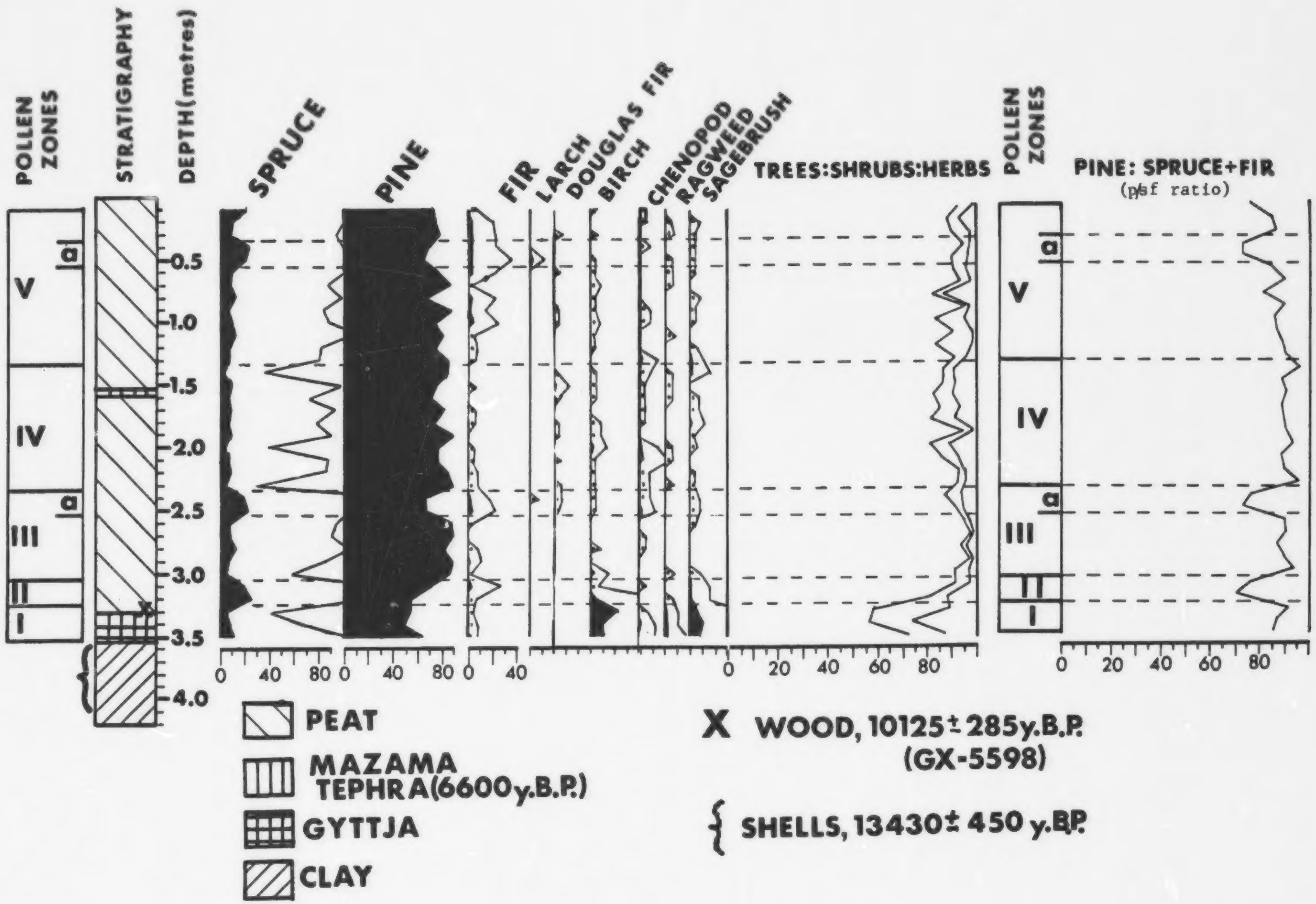


FIGURE 2: Pollen profile, Bog A, upper Elk Valley, British Columbia.

Zone III except for a marked decrease in the pine (66%) to spruce (18-22%) plus fir (2%) ratio. Low percentages of Douglas fir are present.

Zone IV is marked by a constant occurrence of Douglas fir in low percentages. Pine (71-89%) and spruce (4-10%) dominate the vegetation. The p/sf ratio is constantly high. Willow (1-7%) and chenopod (0-2%) percentages increase. There is a slight overall increase in herb percentages.

Zone V is identified by an increase in spruce (9-22%) and fir (0-4%) and a decrease in Douglas fir (*Pseudotsuga menziesii*). The p/sf ratio overall decreases. There is also a decrease in herb percentages. Subzone Va is recognized by an abrupt decrease followed by an increase in the p/sf ratio. This fluctuation was greater in magnitude than the overall trend of p/sf ratio decrease for the zone.

A RECONSTRUCTION OF PAST VEGETATION FOR UPPER ELK VALLEY

Reconstruction of past vegetation was done in part by using the p/sf ratio to infer fire frequency. The effect of fire on the subalpine forest region has been studied by Habeck and Mutch (1973) and Loope and Gruell (1973).

Vegetation in the Elk Valley has been greatly affected by fires. The earliest written record of vegetation in the valley indicates that large tracts had been destroyed by fire (Dawson 1885). A major fire occurred in 1936 in which most of the valley floor and side vegetation was burned. After the fire, Lodgepole pine (*Pinus contorta*) became the dominant tree. This is due to its pioneer characteristic under conditions of disturbance. The climax forest trees for the upper Elk Valley consist of Engelmann spruce (*Picea engelmanni* Parry) and Subalpine fir (*Abies lasiocarpa* [Hook.] Nutt.) (Rowe 1959).

A very high percentage of pine pollen is interpreted in this study to indicate the occurrence of fires that prevent the vegetation from reaching climax forest type. When pine pollen percentages decrease, a reduction in the fires is interpreted to have occurred, allowing vegetation to reach a climax forest type.

The ratio of pine to spruce plus fir pollen counts for each sample is listed in Figure 2. Presumably, a high p/sf ratio, after the original establishment of the vegetation, indicates a subclimax forest type. A lower p/sf ratio indicates a low occurrence of fires. A high p/sf ratio is inferred to indicate the occurrence of major fires.

The bottom of the core (4.2 to 3.6 m), consisting of clay, is characterized by only a few bisaccates and unidentifiable pollen grains, indicating that either there was no vegetation near the bog or that pollen was not preserved. The zone represents a low energy, lacustrine environment in which molluscs and ostracodes were deposited in clay. The gradational contact between the clay and the lower till, combined with lack of pollen in the clay, suggests that the clay layer was formed just after deglaciation.

Pollen Zone I

This pollen assemblage is interpreted to represent a low shrub-herb tundra, which was the first vegetation in the area following deglaciation. The main pioneer assemblage consists of birch and sagebrush with minor amounts of sedge, willow, alder, soapberry, chenopod, ragweed, heath and horsetail. The high percentages of pine and spruce are interpreted to indicate long distance transportation. This has been found in modern tundra environments where the low pollen production of tundra vegetation has the effect of exaggerating the proportions of pollen such as pine, spruce and alder that arrive via long distance transportation (Lichti-Federovich and Ritchie 1968).

Pollen Zone II

Zone II includes the lowest p/sf ratio for the entire profile. The vegetation is dominated by spruce and fir. However, the pine percentage increases through the upper part of the zone, indicating probable movement of pine toward the bog. This zone is interpreted to indicate a replacement of the shrub-herb tundra by spruce and fir before pine developed in the area.

Pollen Zone III

This zone is distinguished by a very high, constant percentage of pine pollen. This is interpreted to indicate the movement of pine into the area. The vegetation consists of pine with some spruce and fir. There are very few shrubs and herbs.

Subzone IIIa is identified by a relatively large decrease in the p/sf ratio. This is inferred to indicate the formation of a climax vegetation of spruce and fir due to a reduction in fires. Sedge and Douglas fir are also present. By assuming a constant sedimentation rate in the peat between the peat/gyttja contact (10,125 ± 285 yr B.P.) and the Mazama tephra layer (6600 yr B.P.), the aproximate maximum and minimum ages of subzone IIIa are 8700 yr B.P. and 8300 yr B.P. respectively.

Pollen Zone IV

The p/sf ratio is constantly high for this zone indicating a subclimax forest due to fires. Douglas fir, shrubs and herbs increased from below indicating continued warming.

Sudworth (1918) has pointed out that the altitudinal limit of Douglas fir is temperature-conditional. At present the closest occurrence of isolated stands of Douglas fir at lower elevations is approximately 20 km south of the study site (personal communication, D. Polster, 1977). The inferred warmer, drier climate of zone IV therefore resulted in the migration of Douglas fir upvalley by at least 20 km. The actual upvalley movement of the main Douglas fir stands was probably greater. Since Douglas fir occurs as a pioneer species following fire in the lower edge of the spruce/fir zone (Habeck and Mutch 1973), the presence of Douglas fir pollen in constant low percentages probably represents an upward altitudinal shift of the lower treeline. The shift is approximately 20 to 25 m.

This pollen zone ends immediately after the deposition of Mazama tephra.

Pollen Zone V

The overall decrease in the p/sf ratio for this interval is interpreted to indicate the vegetation attaining a climax forest cover. This is inferred to indicate fewer fires for this interval. The decrease in Douglas fir indicates a cooler/moister climate.

Subzone Va is identified by a fluctuation in the p/sf ratio which is interpreted to indicate a Late Holocene climax vegetation.

A RECONSTRUCTION OF PAST CLIMATE FOR UPPER ELK VALLEY

Only tentative conclusions regarding the Holocene environement for the upper Elk Valley can be determined by the pollen stratigraphy and vegetation.

The shrub-herb tundra vegetation in pollen Zone I is interpreted to indicate a relatively cool climate following deglaciation. In Zone IV, a warmer/drier climate is inferred to have occurred. The decrease in Douglas fir in Zone V indicates a cooler/moister climate. The relationship between vegetation and climate in Zones II and III is less definite.

There appears to be a possible relationship between the p/sf ratio and climatic changes. There is insufficient data at this point to determine the cause of the relationship, since it is reported that a cooler/moister climate does not necessarily cause a reduction in fires (Habeck and Mutch 1973). In Zone IV, where the presence of Douglas fir is interpreted to indicate a warmer/drier climate, the p/sf ratio is high. In Zone V, where the absence of Douglas fir is interpreted to indicate a cooler/moister climate, the p/sf ratio is low. Therefore a low p/sf ratio may indicate a cooler/moister climate. Subzone IIIa is contradictory to this interpretation. Subalpine larch (*Larix lyallii* Parl.) occurs along the upper altitudinal boundary of the interior subalpine forest region (Rowe 1959). The pollen which was identified as Douglas Fir may therefore be larch. If so, the treeline may have shifted downward during the period covered by this pollen zone.

The sequence of interpreted climatic changes is that Zone I indicates a shrub-herb tundra during a cool climate following deglaciation. The following increase in spruce indicates a vegetation succession and does not have any climatic significance. The large increase in pine in Zone III indicates a warmer climate. Subzone IIIa represents a cooler/moister period. The climate then became warmer/drier during Zone IV in which Douglas fir migrated upvalley to the study site and spruce and fir migrated farther upvalley. Climate then began to deteriorate during Zone V, causing a decrease in Douglas fir and an increase in spruce and fir due to a downward shift of the treeline. Subzone Va indicates a cooling period during the late Holocene.

Since the following interpretations are based on data from only one pollen core, further research is required to more accurately determine the relationship of the vegetation to climatic changes.

RELATIONSHIP OF VEGETATIONAL HISTORY IN ELK VALLEY TO ADJACENT AREAS

Heusser (1956) states that in bog studies in Jasper Park, Alberta the generalized vegetational sequence begins with pine-spruce-fir and succeeds to pine-spruce-Douglas fir, then to pine-spruce-fir and finally to pine-spruce-western hemlock. Although this sequence approximately agrees with that in Elk Valley, there is a definite difference in the pioneer vegetation assemblage following deglaciation.

The Bog A pollen profile indicates a shrub-herb tundra of birch-sagebrush with minor amounts of sedge, willow, alder, soapberry, chenopod, ragweed, heath and horsetail occurring after deglaciation. A similar pioneer assemblage was reported by Ritchie and Hare (1971) in the Mackenzie Delta area. They conclude that the bottom pollen zone, which was bracketed by radiocarbon ages of 12,900 ± 170 yr B.P. (GSC-1321) and 11,500 ± 200 yr B.P. (GSC-1237), indicates that the first vegetation in the area following deglaciation was a low shrub tundra dominated by dwarf birch, with abundant soapberry (*Shepherdia canadensis*) and substantial sagebrush-herb associations. Except for minor differences in pollen percentages and the absence of juniper (*Juniperus*) at the bottom of Bog A, the two assemblages are remarkably alike. Another possibly similar pioneer assemblage was recorded by Lichti-Federovich (1970, 1972) in central Alberta. This assemblage differs from Zone I for Bog A by having a very high poplar pollen percentage and a lower birch percentage. The lack of poplar pollen in Bog A may be due to its known poor preservation characteristics.

MacDonald (1982) recognized two distinct pollen zones in the Morley Flats - Kananaskis Valley area of Alberta. The basal zone, dated to prior to approximately 10,000 yr B.P., is characterized by *Artemisia-Salix-Juniperus* which corresponds to Zone I of this study and is also characterized by pioneer species. Mott and Jackson (1982) demonstrated that a sparse herbaceous tundra-like vegetation dominated the Chalmers Lake area of southwestern Alberta between about 18,400 and prior to 8220 yr B.P. This sequence is similar to Zone I of the Elk Valley sequence. Holloway *et al.* (1981) identified a tundra vegetation sequence between 16,000 and 11,750 yr B.P. at Wabamun Lake, Alberta that is very similar to the Elk Valley sequence.

Therefore, we suggest that following deglaciation a shrub-herb tundra vegetation developed in the upper Elk Valley and that this was a characteristic feature of postglacial pioneer vegetation in southeastern British Columbia and southwestern Alberta. The pioneer assemblage was replaced by a spruce forest. This postglacial forest succession is contrary to that of Heusser (1960) and Hansen (1948, 1949A, 1949B, and 1955). They indicate that lodgepole pine (*Pinus contorta*) was the dominant pioneer tree following deglaciation of the mountains. The record of the earlier shrub-herb tundra which was replaced by the spruce assemblage of Bog A may be due to a more complete pollen record preserved in the gyttja layer than in most bogs.

Harrison (1976) reports that a high percentage of pine pollen from an organic layer dated at 11,900 ± 100 yr B.P. (GSC-2142) and 12,200 ± 160 yr B.P. (GSC-2275) indicates a recently deglaciated area. This locality is only 25 km south of Bog A. If the treeless assemblage zone was missed by Harrison's random pollen sample, then the date for deglaciation in the area could be significantly older than Harrison's 12,000 yr B.P. date.

Pollen Zones III and IV represent the warming trend during the postglacial, which has been called the Altithermal. These pollen zones are similar to the pine-spruce-Douglas fir interval of Heusser (1956). If equal sedimentation rates are assumed for the peat deposits between the dated levels, then this warm period occurred from 9700 to 8700 yr B.P. and 8300 to 5700 yr B.P. This time range is within the interpreted warmer period of pollen Zone L3 of Lichti-Federovich (1970). She determined that the warm period occurred from 9200 yr B.P. to 3500 yr B.P. with a maximum at 5500 to 6000 yr B.P., after which climate gradually deteriorated. The change from subalpine vegetation to grasslands during the Altithermal that occurred in subalpine vegetation of the Livingstone Range, southwestern Alberta (Reeves and Dormaar 1972) did not occur in upper Elk Valley. There was an upward altitudinal shift of the subalpine forest of greater than 20 m. The termination of pollen in Zone IV is approximately the same age as the termination of the Altithermal proposed by Mathewes and Rouse (1975) and Alley (1976). They indicate that the Altithermal ended around the time of the deposition of Mazama tephra. Christensen and Hills (this volume) place the end of the Altithermal at about 5000 yr B.P. in the Waterton Lakes area of southwestern Alberta.

Subzone IIIa, which has an increase in spruce and fir is interpreted to indicate a cooler climate than strata above and below. By assuming equal sedimentation rates, the zone occurred from 8700 yr B.P. to 8300 yr B.P. Reeves and Dormaar (1972) indicate that a major cooling of the climate occurred from 8500 yr B.P. to 8000 yr B.P., during which a downward altitudinal shift in treeline of 600 m occurred. A shift of this magnitude did not occur in the upper Elk Valley, since this would have resulted in treeline below Bog A during this interval. Alley (1972) indicated that a brief cooling period inferred from high spruce and fir percentages occurred below Mazama tephra in the sediment core from Callum Bog, southwestern Alberta, however, the date of this cooler period was not determined.

There is increasing evidence of a late Pleistocene or early Holocene Cordilleran glacial advance in southwestern Alberta. Since radiocarbon dates for the maximum and minimum ages of this advance have not been determined, it is uncertain whether the advance is the Eisenhower Junction (Rutter 1972; Jackson 1977) or Crowfoot (Luckman and Osborn 1979) or both (it may be later proven that the deposits mapped as Eisenhower Junction and Crowfoot are from the same advance). If pollen Subzone IIIa indicates the cooler moister climate which resulted in the Eisenhower Junction and/or Crowfoot advance, then maximum and minimum ages for that advance can be determined from radiocarbon dates on the peat deposits. Rutter (1977) determined that a minor glacial advance, which he called the Deserters Canyon advance, occurred in the Williston Lake area, British Columbia between 9280 ± 220 yr B.P. (GSC-1497) and 7470 ± 150 yr B.P. (GSC-1161). This corresponds to the cooler interval between about 8700 and 8300 yr B.P. in Elk Valley.

Pollen Zone V is assigned to the late postglacial cooler/moister-than present time of Heusser (1956, 1960) in which the vegetation consisted of pine-spruce-fir. The cooler period interpreted from the vegetation in pollen Subzone Va probably relates to one of the late Holocene pollen subzones (KB3a, KB3c and KB3e) of Alley (1976). Alley (1976) interprets the subzones to reflect paleoclimatic changes of minor but regional extent. He tentatively correlates the climatic fluctuations with the stades of the Neoglaciation.

SUMMARY

Sediments from a bog in the upper Elk Valley, British Columbia have yielded a pollen profile extending back at least 13,500 years. The profile is subdivided into five pollen zones. Zone 1 (oldest) is interpreted to represent a *Betula - Artemisia* sagebrush shrub tundra with *Salix, Alnus, Shepherdia*, chenepods, minor heath and *Equisetum*. Zone II, dominated by *Picea* and *Abies* with *Pinus* increasing in the upper levels, represents replacement of the tundra by coniferous forest. Zone III, characterized by high *Pinus*, is interpreted to represent movement of this conifer into the area. A decrease in the pine to spruce plus fir ratios in Subzone 3a is inferred to indicate establishment of climax spruce-fir forest. Zone IV is interpreted to indicate subclimax forest due to fire. Increases in Douglas fir points to continued upslope movement of the vegetation. Zone V (youngest), characterized by an overall decrease in the pine to spruce plus fir ratios and a decrease in Douglas fir, is interpreted to indicate a cooler/moister climate than that of the preceding zone.

ACKNOWLEDGEMENTS

This research was supported by Natural Sciences and Engineering Research Council of Canada Grants in Aid of Research of L.V. Hills and G. Osborn, and by the National Museum of Natural Sciences Climatic Change Project. The many helpful suggestions of G. Osborn and L.E. Jackson, Jr., are greatly appreciated. Tari Forrest typed the manuscript.

REFERENCES

Alley, N.F. 1972. The Quaternary history of part of the Rocky Mountains, Foothills and Plains and Western Porcupine Hills, southwestern Alberta. PhD. thesis, University of Calgary, Calgary, Alberta. 201 pp.

_______. 1976. The palynology and paleoclimatic significance of a dated core of Holocene peat, Okanagan Valley, southern British Columbia. Canadian Journal of Earth Sciences 13:1131-1144.

Bujak, C.A. 1974. Recent palynology of Goat Lake and Lost Lake, Waterton Lakes National Park. M.Sc. thesis, University of Calgary, Calgary, Alberta. 60 pp.

Christensen, O.A., and L.V. Hills. 1971. Palynologic and paleoclimatic interpretation of recent sediments, Waterton Lakes National Park, Alberta. Report to National and Historic Parks Branch, Department of Indian Affairs and Northern Development. 15 pp.

Dawson, G.M. 1885. Preliminary report on the physical and geological features of that portion of the Rocky Mountains between latitudes 49° and 51°30'N. Geological Survey of Canada, Report of Progress for 1881-83-84, Part C. 169 pp.

Driver, J.C. 1978. Holocene man and environments in the Crowsnest Pass, Alberta. PhD. thesis, University of Calgary, Calgary, Alberta. 230 pp.

Faegri, K., and I. Iverson. 1965. Field techniques. In: Handbook of Paleontological Techniques. Edited by: B. Kummel and D. Raup. W.H. Freeman and Company, San Francisco. pp. 482-493.

Fergusson, A.J. 1978. Late Quaternary geology of the upper Elk Valley, British Columbia. M.Sc. thesis, University of Calgary, Calgary, Alberta. 118 pp.

Fergusson, A.J., and G. Osborn. 1981. Minimum age of deglaciation of upper Elk Valley, British Columbia. Canadian Journal of Earth Sciences 18:1635-1636.

Gray, J. 1965. Extraction techniques. In: Handbook of Paleontological Techniques. Edited by: B. Kummel and D. Raup. W.H. Freeman and Company, San Francisco. pp. 530-587.

Habeck, J.R., and R.W. Mutch. 1973. Fire-dependent forests in the northern Rocky Mountains. Quaternary Research 3:408-421.

Hansen, H.P. 1948. Postglacial forest of the Glacier National Park region. Ecology 23:146-152.

______. 1949A Postglacial forests in south central Alberta, Canada. American Journal of Botany 36:54-65.

______. 1949B Postglacial forests in west central Alberta, Canada. Bulletin of the Torrey Botanical Club 76:278-289.

Hansen, H.P. 1955. Postglacial forests in southcentral and central British Columbia. American Journal of Science 253:640-658.

Harrison, J.E. 1976. Dated organic material below Mazama(?) Tephra: Elk Valley, British Columbia. Geological Survey of Canada Paper 76-1C:169-170.

Heusser, C.J. 1956. Postglacial environments in the Canadian Rocky Mountains. Ecological Monographs 26:253-302.

______. 1960. Late-Pleistocene environents of North Pacific North America. American Geographical Society Special Publication 35:1-308.

Holloway, R.G., V.M. Bryant, Jr., and S. Valastro. 1981. A 16,000 year pollen record from Lake Wabamun, Alberta, Canada. Palynology 5:195-207.

Jackson, L.E. 1977. Quaternary stratigraphy and terrain inventory of the Alberta portion of the Kananaskis Lakes 1:250,000 sheet (82-J). PhD. thesis, University of Calgary, Calgary, Alberta. 480 pp.

Kapp, R.O. 1969. Pollen and spores. Wm. C. Brown Co., Dubuque, Iowa. 249 pp.

Lichti-Federovich, S. 1970. The pollen stratigraphy of a dated section of late Pleistocene lake sediment from central Alberta. Canadian Journal of Earth Sciences 7:938-945.

______. 1972. Pollen stratigraphy of a sediment core from Alpen Siding, Alberta. Geological Survey of Canada Paper 72-1B:113-115.

Lichti-Federovich, S. and J.C. Ritchie. 1968. Recent pollen assemblages from the western interior of Canada. Review of Palaeobotany and Palynology 7:297-344.

Loope, L.L., and G.E. Gruell. 1973. The ecological role of fire in the Jackson Hole area, northwestern Wyoming. Quaternary Research 3:425-443.

Luckman, B.H., and G.D. Osborn. 1979. Holocene glacier fluctuations in the middle Canadian Rocky Mountains. Quaternary Research 11:52-77.

MacDonald, G.M. 1982. Late Quaternary paleoenvironments of the Morley Flats and Kananaskis Valley of southwestern Alberta. Canadian Journal of Earth Sciences 18:23-35.

Mathewes, R.W. 1973. Paleocology of postglacial sediments in the Fraser Lowland region of British Columbia. Ph.D. thesis, University of British Columbia, Vancouver. 77 pp.

Matthewes, R., and G.E. Rouse. 1975. Palynology and paleocology of postglacial sediments from the Lower Fraser River Canyon of British Columbia. Canadian Journal of Earth Sciences 12:745-756.

Mott, R.J., and L.E. Jackson, Jr. 1982. An 18,000-year palynological record from the southern Alberta segment of the classical Wisconsin "Ice Free" Corridor. Canadian Journal of Earth Sciences 19:504-513.

Reeves, B.O.K., and J.F. Dormaar. 1972. A partial Holocene pedological and archeological record from the southern Alberta Rocky Mountains. Arctic and Alpine Research 4:325-336.

Reubin, M. and Alexander, C. 1960. United States Geological Survey radiocarbon dates. Radiocarbon 2:129-185.

Ritchie, J.C., and F.K. Hare. 1971. Late-Quaternary vegetation and climate near the arctic treeline of northwestern North America. Quaternary Research 1:331-342.

Rowe, J.S. 1959. Forest regions of Canada. Forestry Branch Bulletin 123, Department of Northern Affairs and Natural Resources, Ottawa.

Rutter, N.W. 1972. Geomorphology and multiple glaciation in the area of Banff, Alberta. Geological Survey of Canada Bulletin 206:1-54.

Rutter, N.W. 1977. Multiple glaciation in the area of Williston Lake, British Columbia. Geological Survey of Canada Bulletin 273:1-31.

Sudworth, G.B. 1918. Miscellaneous conifers of the Rocky Mountain region. United States Department of Agriculture Bulletin 680:1-45.

Weir, G.H., and E.L. Thurston. 1976. Scanning electron microprobe identification of fossil Pinaceae pollen to species by surface morphology. Palynology 1:157-165.

PART 3: HOLOCENE PALYNOLOGY OF CROWSNEST LAKE, ALBERTA, WITH COMMENTS ON HOLOCENE PALEOENVIRONMENTS OF THE SOUTHERN ALBERTA ROCKIES AND SURROUNDING AREAS

J.C. Driver, L.V. Hills and B.O.K. Reeves

INTRODUCTION

Late Pleistocene and Holocene palynological studies have recently received renewed attention with the location of relatively long records at Wabamun Lake, Alberta (Holloway *et al.* 1981), Chalmer's bog (Mott and Jackson 1982), Kananaskis Valley, Alberta (MacDonald 1982) and Elk Valley, British Columbia (Fergusson and Hills, this volume). As part of paleoenvironmental reconstructions for southern Alberta, a number of lake cores were taken and studied palynologically (Alley 1972; Christensen and Hills, this volume; Fergusson and Hills, this volume; Bujak 1974).

During February 1977, a 6.4 m core of sediment was taken from the eastern end of Crowsnest Lake (see Frontispiece at beginning of this series of papers). The base of the deposits in the lake was not reached because at about 6 m deposits became hard, and were impenetrable for a hand-held Livingstone coring device.

The core was split longitudinally, samples of approximately 2 cc were taken from the interior of one half at regular intervals and the balance was screened for macrofossils. The other half of the core was frozen and stored.

Pollen samples were initially processed in a 10% HCl solution to remove carbonates, and were then left in 52% HF solution for 24 hours to remove silicates. The residue was treated with Schultz oxidizing solution (HNO_3 + $KClO_3$) for 10 minutes to remove extraneous organic detritus. Ten percent K_2CO_3 was added to remove humic acids. Finally the pollen-bearing sediment was stained with Saffronin O, and slides were prepared.

Before the zonation is discussed in detail, it should be noted that pollen preserved in sediments at the east end of Crowsnest Lake may have originated from a variety of communities. Prevailing winds in the Crowsnest area are westerly, and thus much of the pollen probably originated west of the sampling site.

Lack of "exotic" pollen, notably mountain and western hemlock (*Tsuga mertensiana* and *Tsuga heterophylla*), indicates that little pollen from the Columbia forest zone of the Rocky Mountain Trench to the west penetrated the Crowsnest Pass. Although some pollen may have originated from the Elk River Valley (Frontispiece, shaded area) or the British

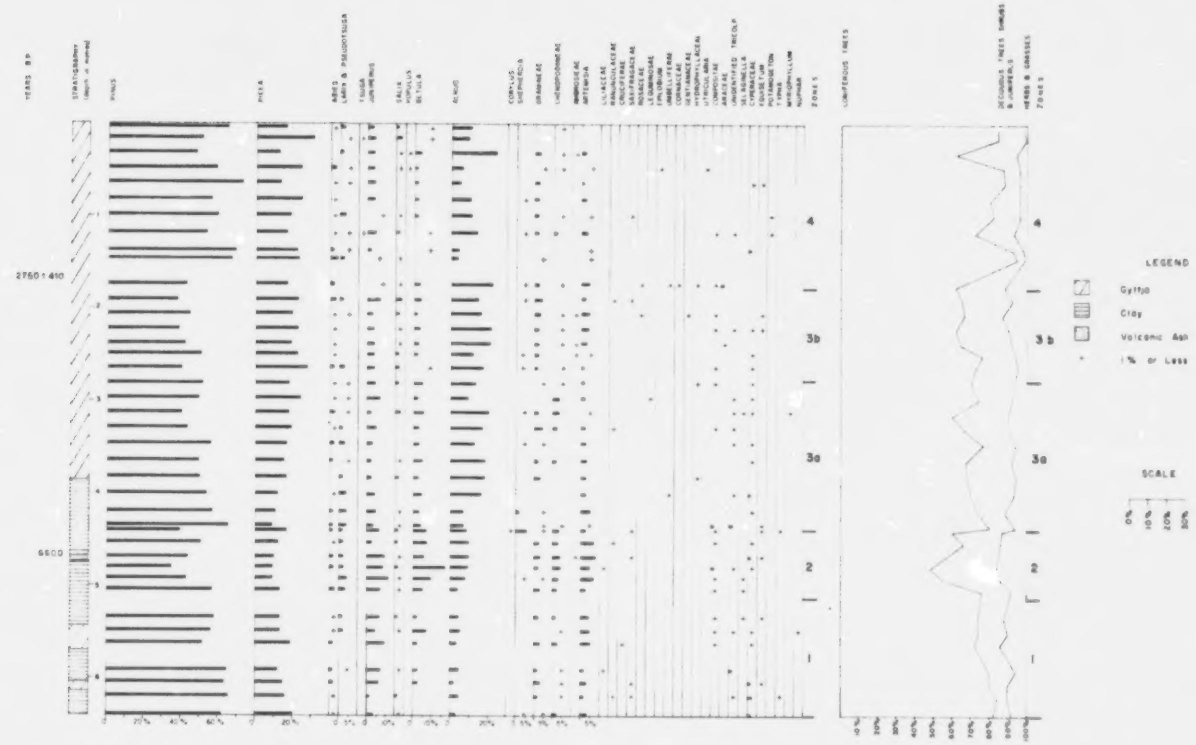


FIGURE 1: *Relative percentages of pollen, Crowsnest Lake, Alberta (cored 1977).*

Columbia section of the Crowsnest Pass, low values for larch (*Larix* sp.) suggest that only small amounts of pollen were derived from that region. While some pine (*Pinus* spp.) pollen may have travelled long distances before being deposited in Crowsnest, the majority of pollen from other species is likely to have been produced within or close to Crowsnest Pass, particularly to the west of the sampling site. Some areas west of the site today support grasslands and subalpine forest. Montane forest is poorly represented, and alpine zones, while covering a considerable area, support little vegetation.

Dating of any part of the Crowsnest Lake core is difficult, with the exception of the Mazama tephra level, which dates to approximately 6600 yr B.P., and the single radiocarbon dated sample at 1.75 m dating to 2750 ± 410 yr B.P. (RL - 861). A date of approximately 9000 yr B.P. for the base of the core is estimated. The 7000 yr B.P. date is also an estimate.

POLLEN ZONES

The Crowsnest Lake core (Figure 1) can be subdivided into four zones as follows:

Zone 1: 6.4 to 5.1 m (ca. 9000 to ca 7000? yr B.P.)

Characteristics of Zone 1 are moderate values for total coniferous tree pollen (*ca* 80%), low values for deciduous trees, shrubs and *Juniperus* (*ca* 10%), and moderate values for herbs and grasses (*ca* 10%). Of the conifers, pine and spruce (*Pinus* and *Picea*) are the most important, and this situation pertains throughout the core. Their presence suggests a coniferous forest, similar to modern subalpine forest. The forest was well established by 9000 yr B.P. Compared with the top of the core, which represents modern vegetation, *Picea* percentages in Zone 1 are somewhat lower indicating that the lower treeline of the subalpine forest was further upslope from the lake than it is today, and thus that the climate was slightly warmer and drier than modern conditions.

Zone 1 values for alder (*Alnus*) are low when compared with modern pollen frequencies. *Alnus* is problematic because it grows in wet areas, and thus local shifts in habitat around a lake may affect its occurrence. Although short-term fluctuations are probably of little climatic significance, long-term trends may be important. Low values for *Alnus* in Zone 1 may reflect climate, and could show that lake level tended to be stable and high at this time.

Indicators of grassland, notably Gramineae, Chenopodiineae, and *Artemisia* (grasses, chenopods, and wormwood or sagebrush) are moderately well represented during Zone I, particularly towards the upper part of the zone. Their values are higher than for samples at the top of the core, suggesting that grassland was more extensive than today. *Artemisia* is usually characteristic of open grassland conditions in the Crowsnest area today, but is more commonly found in the Foothills and Plains east of the Pass. Although some species of *Artemisia* are found in alpine grasslands, most are found below the lower treeline, and it is likely that the *Artemisia* found in the Crowsnest core represents one or more of these latter species.

Pollen frequencies in Zone 1 suggest that climate was warmer and drier than today. Lower treelines were further upslope from the sampling site than today, as shown by lower values of *Picea*. This implies an increased area of valley grasslands, which is further supported by high *Artemisia* values and moderate values for Gramineae. The presence of montane forest is difficult to document for any zone in the Crowsnest core. Pollen of both *Larix* (larch) and *Pseudotsuga* (Douglas fir) is similar, but can be separated on the basis of size (Kapp 1969, pp. 64-65). Measurements on the pollen from Crowsnest are difficult to take because the large inaperturate pollen grains typical of both genera are often ruptured. For this reason, *Larix* and *Pseudotsuga* are grouped together on Figure 1. Data from Alley (1972) suggest that montane forest was not present in southwestern Alberta at the time. If so, the forest/grassland transition in Crowsnest during this period may have been dominated by lodgepole pine (*Pinus contorta*). This interpretation is in keeping with findings of Christensen and Hills (this volume) and Fergusson and Hills (this volume).

Zone 2: 5.1 to 4.3 m (ca. 7000 yr B.P. to ca. 5000 yr B.P.)

The change to Zone 2 is defined by a sudden decline in conifer pollen. Certain aspects of Zone 2, notably high values for *Artemisia*, were initiated in the upper part of Zone 1. This trend is accentuated in Zone 2, where high values for grasses and herbs appear.

Pinus and *Picea*, both decline and reach their lowest values in the core in Zone 2. This is probably due to an upslope movement of lower treeline away from Crowsnest Lake, and, considering the long dispersal distances possible for *Pinus* pollen, may also reflect a regional decline in the amount of coniferous forest. Grasslands within the Crowsnest Pass were more extensive than during Zone 1, and more extensive than today. A warm, dry climate would favour expansion of grasslands into areas where forest could not reproduce because of low ground moisture.

Increases in birch (*Betula*) and alder (*Alnus*) are more difficult to explain. Both genera are often associated with wet conditions, which would appear to contradict the above interpretation of a relatively warm, dry climate. However, it is possible to explain the presence of both in the context of such an environment. The *Betula* pollen probably belongs to a small, shrubby species - *Betula glandulosa*. The small size of the pollen grains precludes either *Betula papyrifera* or *B. occidentalis*. *Betula glandulosa* is not common in Crowsnest Pass today, but it is found in other areas of the Rockies. It occurs abundantly on well-drained, stony soils of the Bow River and Kananaskis Valley west of Calgary and in the Upper Elk Valley to the west of the study area. Here it is in association with grasslands and montane forest. Possibly *B. glandulosa* began to colonize certain areas of the Crowsnest Pass, especially terraces composed of glacial outwash gravels, as subalpine forest retreated.

Alnus pollen could have been derived from a number of species, however *A. crispa* is the most common species in Crowsnest Pass today. It grows along streams, and generally prefers a wet habitat. It may favour places where seasonal inundations take place. Possibly the increase in *Alnus* during Zone 2 is related to water levels that fluctuate seasonally, perhaps due to lack of summer rainfall.

Thus, Zone 2 almost certainly represents a period which was relatively warmer and drier than both Zone 1 and the modern climate. As in Zone 1, the presence of montane forest cannot be positively documented, although *Betula glandulosa* is found today in association with *Pseudotsuga menziesii*. Large inaperturate pollen of the *Larix/Pseudotsuga* type is more frequent in Zone 2, and it is tempting to believe that this represents the establishment of montane forest zone. This is certainly the period when montane forest should appear according to Alley's (1972) nearby palynological sequence. The degree to which vegetation differs from the modern situation is probably greater during Zone 2 than in any other part of the core. Although quantitative estimates are difficult to make, the following qualitative changes can be inferred to have taken place in the distribution of vegetation zones at this time:

(a) Expansion of grasslands in creek valleys entering the main Pass from the north, particularly Gold Creek, Pelletier Creek and Blairmore Creek, all of which contain some valley grassland under modern conditions.

(b) Expansion of microtopographic/climatic areas of grassland within the subalpine forest, particularly in the northern tributary valleys.

(c) Possible establishment of microtopographic/climatic grassland in tributary valleys where this feature is lacking today.

(d) Altitudinal expansion of valley grasslands, particularly to the north of the Pass in basins west of Coleman, west of Blairmore, and around Bellevue.

(e) Spread of montane forest into areas now occupied by subalpine forest. Montane forest may have occupied all slopes of the Livingstone Range up to 2000 m (the highest modern recorded occurrence of Douglas fir), northern slopes of the Pass and lower reaches of northern tributary valleys, and southern terraces of the Pass.

(f) A reduction in area of *Picea* subalpine forest, due to its lower tolerance of dry conditions. Lower treelines probably gained altitude, while upper treelines may not have gained significantly in all areas because topographic factors would still control upper treeline to some extent.

(g) A possible reduction in the area of alpine vegetation in those areas where subalpine forest upper treelines could rise in altitude and invade alpine meadows and grasslands.

This interval closely corresponds to Zone II at Linnet Lake (Christensen and Hills, this volume), and is in keeping with the continued upslope migration of vegetation (Alley 1972; Fergusson and Hills, this volume; MacDonald 1982; Mott and Jackson 1982).

Zone 3: 4.3 to 1.75 m (ca. 5000 yr. B.P. to ca. 3000 yr. B.P.)

Zone 3 is characterized by values for coniferous pollen between those of Zone 2 and modern values, high amounts of *Alnus*, and moderate values for herbs and grasses. Zone 3 is subdivided into Subzones 3a and 3b, the latter differing from the former by increased percentages of Gramineae and *Artemisia*

Subzone 3a represents cooler and wetter conditions than the preceding Zone 2. Conifers, particularly *Pinus* and *Picea*, are more frequent, suggesting that the lower treeline was advancing down the valley sides, taking advantage of increased soil moisture. Both *Betula* and *Juniperus* decrease from Zone 2. *Juniperus* probably represents creeping juniper (*Juniperus horizontalis*) which is found today on open, stony areas. *Juniperus communis* grows in sheltered understory conditions, and therefore its pollen is probably not picked up and wind transported. Reduction of it and *Betula* suggests a decrease in open areas around the lake. Similarly, grassland pollen of Gramineae, Chenopodiineae, and *Artemisia* also decreases. *Alnus*, which may represent wet conditions, increases in Subzone 3a. Most palynological evidence suggests that at around 5000 yr B.P. there was an abrupt change in climate, with reduced temperatures and increased precipitation. When compared to today, however, it was still slightly warmer and drier. Throughout the Crowsnest Pass, grassland and montane forest were reduced in area, and grassland probably disappeared from many of the tributary valleys that it had invaded during Zone 2. Subalpine forest increased in area. Greater values for *Alnus* suggest that previously dry areas were trapping increased moisture and becoming ponds or swamps with local growth of *Alnus*. Subzone 3a appears to correspond to Zone III of Christensen and Hills (this volume).

Although many features of Subzone 3b are similar to Subzone 3a, subzone 3b time was probably warmer and drier. Increased values of Gramineae and *Artemisia* document an expansion of valley grasslands. However, coniferous pollen during this period is more dominated by *Picea* than in earlier periods. This suggests that subalpine forest did not retreat far from the sampling site. Perhaps the reduced values of *Pinus* encountered in this zone may indicate that grassland was moving into lower areas of the subalpine forest that are today dominated by *Pinus contorta*. It is also possible that the grassland invasion only took place on south-facing slopes of Crowsnest Pass. This would maintain a thick growth of subalpine forest on north-facing slopes such as those to the south of Crowsnest Lake at the present time. Forest of this type could thus contribute *Picea* pollen to the sampling site, while forests on south-facing slopes, which tend to contain more *Pinus*, would contribute less to the pollen sample as they were reduced in area.

The Subzone 3b warm period is not marked by major increases in *Juniperus* or *Betula*, such as occurred in Zone 2. This suggests less expansion of open areas, and also little expansion of montane forest. The latter case is supported by low values for *Larix/Pseudotsuga*. Apparently, Subzone 3b was a minor warm period in the Crowsnest sequence. Probably it was shorter than Zone 2 time, and its effects were less widespread, being confined largely to the more sensitive south-facing slopes. Subzone 3b appears to correspond closely to Zone II of Christensen and Hills (this volume). These fluctuations were not recorded in other studies, probably reflecting the favourable position of both Linnet and Crowsnest lakes to a vegetation zonation boundary.

Zone 4: 1.75 to 0 m (ca. 3000 yr B.P. to the Present)

Zone 4 seems to represent the coldest and wettest conditions for the entire post-9000 yr B.P. sequence. Coniferous pollen is very common, reaching 95% in one sample. Both spruce and pine values are high, showing that coniferous forest increased in area. Herb and grass pollen is exceptionally low during Zone 4; in one sample none was found. Slightly lower values for deciduous trees and shrubs are difficult to interpret, unless one assumes that coniferous forest was so widespread that it masked them.

Although Zone 4 represents colder and wetter conditions than preceding zones, it may contain two cold periods, separated by a slightly warmer interval. Cold periods, in which percentages of herbs and grasses are 1% or less, occur at about 1.5 m and 0.1 m.

SUMMARY: CROWSNEST PASS VEGETATION AND CLIMATE FROM 9000 YR B.P. TO THE PRESENT

The date for the establishment of vegetation zones similar to those of today in Crowsnest Pass is not known, but some were certainly present at 9000 yr B.P. Assuming a minimum 9,000 yr B.P. date for the base of the Crowsnest Lake core, grasslands in the valley were more extensive than today between 9000 and 7000 yr. B.P. Above the grasslands was a subalpine forest, dominated by *Pinus* and *Picea* - although *Larix* and *Abies* may have been important in the upper parts of the forest. It seems unlikely that montane forest was established at this time. It can be inferred that alpine vegetation was also present.

Starting within the 9000 to 7000 yr B.P. period, grasslands expanded, culminating in a considerable expansion of the zone during the period 7000 to 5000 yr B.P. At this time, subalpine forest retreated from the valley bottom. At the same time, montane forest (probably a complex association of grassland, birch and Douglas fir) also expanded in area.

At the east end of the Pass, montane forest and grassland may have completely replaced subalpine forest, but, at the west end, subalpine forest was still present on slopes above the expanded montane forest.

At about 5000 yr B.P., a sudden change in climate ameliorated conditions for subalpine forest species, which recolonized areas of montane forest and grassland. Vegetation boundaries were slightly higher in elevation than today.

A warm period from perhaps 4000 to 3000 yr B.P. allowed grasslands to expand slightly, probably at the expense of subalpine forest on south-facing slopes. Grasslands, however, did not reach the extent recorded during the 7000 to 5000 B.P. interval (Zone 2).

From 3000 yr B.P. to the present, vegetation boundaries in Crowsnest Pass were approximately the same as today, but two periods of pronounced cold and increased moisture occurred, the first just after 3000 yr B.P., and the second a few centuries before present.

COMPARISON OF THE CROWSNEST PASS FOSSIL POLLEN SEQUENCE WITH DATA FROM THE ALBERTA ROCKIES

Late Glacial and Early Postglacial to ca. 9000 yr B.P.

Although there is no palynological evidence from the Crowsnest Pass for this period, people had occupied the area by 10,000 yr B.P., and evidence of earlier occupations may be found (Reeves 1975A). An item of regional palaeoecological interest is a buried soil from Bellevue, which dates to approximately 10,000 yr B.P. (Reeves 1975A). Reeves has suggested that the soil was formed under an upper treeline forest, and that upper treeline was 1000 m lower than it is today. This information can be examined in the light of other evidence from the Alberta Rockies.

Data presented by Alley (1972), Fergusson and Hills (this volume), Holloway *et al.* (1981), MacDonald (1982), and Mott and Jackson (1982) suggest that regional vegetation was characterized by a dwarf birch shrub tundra period to *ca.* 10,000 years ago.

The Late Wisconsin glacial sequence in the Alberta mountains is poorly known, nor is it dated at many localities. Alley (1972) has suggested that the Crowsnest Pass remained unglaciated since Classical Wisconsin time. At Waterton Lakes National Park, Stalker and Harrison (1977) have inferred a Classical Wisconsin advance. For the same area, Reeves (1972B, 1974C, 1975B) has proposed a glacial sequence based upon terminology from the United States Rockies (Richmond 1965). Reeves' sequences include three glaciations: Pinedale I (about 20,000 yr B.P.), Pinedale II (unknown date), and Pinedale III (11,500 to 10,500 yr B.P.). Reeves (1975B) suggests that the postglacial began at 10,500 yr B.P. in the Waterton area, and that, following this, a warming trend occurred which lasted until 8500 yr B.P. During this warm period treelines were higher than today.

Possible palynological evidence for this period comes from an undated core collected from Callum Bog in the Foothills northwest of the Livingstone Range (Alley 1972). The lower zone of the pollen sequence shows that glacial vegetation, possibly dating between 20,000 and 11,000 yr B.P., was succeeded at 11,000 to 10,000 yr B.P. by a warming trend in which forests were established. The latter date is estimated on the basis of similar developments at Yellowstone National Park (Baker 1970; Waddington and Wright 1974) and the Canadian Plains - Saskatchewan and Manitoba (Ritchie 1976) and Alberta (Holloway *et al.* 1981; Mott and Jackson 1982). In both areas, forests were established by 11,000 yr B.P.

North of Crowsnest Pass, many glacial sequences have been described (Harris and Waters 1977), but few have been dated. The final Wisconsin advances in the Bow River Valley are complex. The final major advances are termed Canmore and Eisenhower Junction (Rutter 1972). The latter may correspond to Reeves' Pinedale III in age (Reeves 1974B). This interpretation was supported by dates obtained on Bow River Valley terraces at the eastern edge of the Foothills. However, these terraces and contained gravel deposits are now interpreted as paraglacial gravels (Jackson *et al.* 1982), and hence have no direct glacial origin.

9000 to 7000 yr B.P.

This period in Crowsnest corresponds with Zone 1 of the pollen sequence. It is thought to have been slightly warmer and drier than today, particularly from 8000 to 7000 yr B.P., when the grassland expansion, which peaks in the Altithermal, began.

A soil at archaeological site DjPp-3 on Crowsnest Lake described as an "Alpine Eutric Brunisol (?)" (Reeves 1974A), is dated by typological association of projectile points between 9000 and 8000 yr B.P. On the basis of the thick Ah horizon, the soil is thought to have formed in the "subalpine ecotone" (Reeves 1974B). Reeves suggests that the upper treeline in the Crowsnest Pass was depressed 700 m below modern elevations then, and that much of the study area was under alpine vegetation.

Further pedological evidence comes from an archaeological site at the Gap, where the Oldman River cuts through the Livingstone Range 40 km north of Crowsnest. The Gap is at about the same elevation as the Crowsnest Pass. Excavation of an archaeological site, situated 30 m in elevation above the modern lower timberline, revealed a series of occupations and buried soils. The lowest soil at the Gap is a Degraded Alpine Eutric Brunisol, considered indicative of upper treeline conditions (Reeves and Dormaar 1972). As upper treeline is well above the site today, Reeves and Dormaar suggest that the soil can be accounted for by postulating a treeline which was 600 m lower than today at the time of soil formation. Dates of 8000 ± 150 yr B.P. and 9520 ± 240 yr B.P. have been obtained from the soil. The former date is preferred by Reeves and Dormaar, but they note that the latter date could also be acceptable, as the soil may have been forming for a long time. Reeves and Dormaar did not mean that the soil was a Degraded Alpine Eutric Brunisol for 1500 years. Rather, soil formation may have occurred at the site for an extended period, the distinctive profile developing between 8500 and 8000 yr B.P.

Palynological data from Callum Bog shows that the warming trend from about 11,000 yr B.P. to 6600 yr B.P. was interrupted by a short period when spruce and willow pollen increased, possibly indicating cooler, wetter conditions (Alley 1972). The deposits also change at this time from clays and silts to sands, also indicative of moister conditions. The exact period is undated. Palynological evidence of Fergusson and Hills (this volume), Holloway *et al.* (1981), MacDonald (1982) and Mott and Jackson (1982) indicates that this was, in general, a period of climatic amelioration with no evidence of cooling of climate that would result in lowering of treeline by 600 m.

Waterton Lakes National Park yields evidence for cold conditions at about this time, based upon the presence of cryoturbation at the archaeological site DgPm - 1, dated to 8190 ± 260 yr B.P. (Reeves 1972A,B). Reeves (1972B, 1975B) has equated this with a glacial advance at Waterton termed Lone Creek. Buried soils formed under cool-wet conditions dating to this interval have been found associated with the earliest occupation levels which date *ca* 8000-8500 yr B.P. in two other sites in Waterton Lakes National Park DgPl-1 and DgPl-4 (Reeves 1972B). However, Luckman and Osborn (1979) were unable to locate and hence confirm these advances.

Evidence for glacial activity in other areas of the Alberta Rockies comes from Banff National Park, where Harris and Howell (1977) described two moraines formed by glaciation which took place after the Eisenhower Junction advance (tentatively dated at 11,500 to

10,500 yr B.P.), but before the deposition of Mazama volcanic ash at 6600 yr B.P. The two events, Lake Louise I and II, extended between 4 and 15 km beyond modern glaciers and modern cirques. Lake Louise I was thought to date to a period before 8800 yr B.P., based upon estimated dates for the formation of lakes behind Lake Louise I moraines. Assuming that this interpretation is correct, and assuming that the Eisenhower Junction advance can be equated with the period 11,500 to 10,500 yr B.P., then Lake Louise I must be a hitherto unrecognized event in the Alberta Rockies. Lake Louise II is correlated by Harris and Howell (1977) with the Lone Creek Advance described by Reeves (1972A) for Waterton Lakes National Park. This correlation is not based upon direct dating. Luckman *et al.* (1978) argue that the evidence presented by Harris and Howell (1977) is equivocal and that several of the features mapped as moraines could equally be landforms created during Wisconsinan deglaciation.

Pre-6600 yr B.P. glacial activity has been noted by Osborn and Duford (1976) for the same area. However, they claim that moraines in the period 10,000 to 6600 yr B.P. are very rare, because most moraines of that period were obliterated by Neoglacial advances in the last few centuries. Pre-6600 yr B.P. moraines described by Osborn and Duford do not extend more than 1 km from modern glaciers or modern cirques. They suggest that the most extensive ice advances in the Alberta Rockies during the Holocene were those of the Neoglacial, and that earlier events must have been minor.

From within Crowsnest Pass, a buried soil at Bellevue complicates the picture. A grassland soil at Bellevue has been dated to 8020 ± 200 yr B.P. (RL - 448) (Reeves 1975A). Bellevue is within the modern grassland zone, thus no drastic lowering of treelines can be seen here. While this may produce problems of interpretation when compared with soils at the Gap and at Crowsnest Lake, the Bellevue location is ecologically different, and may never have been occupied by forests as it lies on the north-exposed side of the valley, while the other two sites are on north-facing slopes.

West of Crowsnest Pass, evidence from the Elk River Valley shows that vegetation zones remained nearly unchanged in postglacial time. Study of macrofossils from a bog at 1800 m a.s.l. has shown that *Picea* macrofossils occur throughout the postglacial, and the upper treeline cannot have dropped below 1800 m at any time (Fergusson and Hills, this volume). Treeline limits for both Crowsnest and the Gap have been suggested as 600 to 700 m lower than today, indicating an upper treeline of about 1400 m at 8500 to 8000 yr B.P. (Reeves 1974A, 1974B; Reeves and Dormaar 1972). It is difficult to conceive of a climatic situation which would depress treelines a minimum of 400 m lower in Crowsnest than in the Elk River Valley. Moreover, palynological evidence from the Elk Valley bog suggests that forest levels were at least as high in elevation as today (Fergusson and Hills, this volume). If so, taking both sets of data at face value, the difference in elevation of upper treelines would have been 700 m.

Thus, there is contradictory evidence for the period 8500 to 8000 yr B.P. The palynological evidence presented above and the work of Halloway *et al.* (1981) and Mott and Jackson (1982) indicate that this was a period of general warming and that no major glacial event took place.

Interpretation of soils also poses a problem for this time interval. The soils from Crowsnest Lake and the Gap suggest that vegetation unlike that of today was present at the sites. Conversely, the soil from Bellevue, which dates to the same period, does not differ from the soil which could be expected there today under natural conditions. If we accept the evidence of both soils at Crowsnest, we must assume that the upper treeline was at 1400 m, but that the lower treeline was above 1300 m at Bellevue. This would mean that coniferous forest would have been confined to a thin strip between extensive alpine vegetation and valley grasslands. Such different environmental conditions would certainly register in the Crowsnest palynological core. Both Crowsnest Lake and the Gap are nearer to lower treeline than Bellevue. If the soils at both sites are interpreted merely as forest soils, perhaps a small lowering of lower treeline could have produced different soils, while leaving the Bellevue soil unaffected. The latter solution would be compatible with the palynological evidence.

The Altithermal in the Alberta Rockies

Zone 2 in the Crowsnest core (Figure 2) is considered to represent the widely recognized Altithermal warm period. The dates of 7000 to 5000 yr B.P. for its duration are estimates based upon sedimentation rates within the core.

The onset of the Altithermal is difficult to define for the Alberta Rockies, partly because few studies are available, and partly because the postglacial warming trend begins at about 10,000 yr B.P., after Pinedale III times (Reeves 1974B, 1975C). As noted in the discussion of Zone I of the Crowsnest core, Gramineae or *Artemisia* increase before the lower boundary of Zone 2. Therefore, depending on how "Altithermal" is defined, the period may have begun prior to 7000 yr B.P.

Most data from the Alberta Rockies support an Altithermal warm period. Only one glacial event has been proposed for this period, at Castle River south of Crowsnest Pass, where outwash deposits associated with a terminal moraine were dated to 6200 yr B.P. (Stalker 1969). This event is unique for the Northern Rockies, and is not compatible with the palynological evidence: possibly the dated bones and outwash at the site were not directly associated with the moraine (Luckman and Osborn 1979; Shaw 1972).

Alley's (1972) palynological sequence from Callum Bog shows that the warm trend, which began in postglacial time, culminated just before 6600 yr B.P. when *Pseudotsuga*, Gramineae, Chenopodiineae, and Compositae were more common than at present. *Opuntia* (cactus) pollen was also found in the core. Cactus today grows at least 40 km southeast of Callum Bog.

Altithermal soils from the Gap site have been identified as Orthic Regosols (Reeves and Dormaar 1972). Two Ah horizons, separated by Mazama Ash, have been dated to 6720 ± 140 yr B.P. and 6060 ± 140 yr B.P. The soils are interpreted as having developed under grassland conditions. As the site now lies within a forest, treelines must have been at least 30 m higher than they are today.

A palynological sequence from Linnet Lake, Waterton Lakes National Park, represents conditions in the later part of the Altithermal (Christensen and Hills, this volume). At Linnet Lake, the base of the deposits was found to be Mazama tephra (6600 yr B.P.). Pollen immediately above the tephra shows a period warmer than today, with higher values of *Populus*, *Ambrosia*, *Artemisia*, Gramineae, and Chenopodiineae. The local forest was dominated by *Pinus contorta*. The end of the period has been dated to about 5000 yr B.P. Thus, in Waterton, 6600 to 5000 yr B.P., the Altithermal resulted in an increase in the area of grasslands and in a shift up-valley of subalpine forest. The interpretation for Zone 2 at Crowsnest is similar to this.

End of the Altithermal to 3000 yr B.P.

In the Crowsnest Lake palynological sequence, Subzone 3a has been described as a return to cool conditions, with a reduction in area of valley grasslands and montane forest, and an expansion of subalpine forest. This corresponds to Zone II at Linnet Lake and may correlate with a period of rock glacier activity between the Crowfoot and Cavell advances described by Luckman and Osborn (1979).

Reeves (1972A) has postulated a minor glaciation at Waterton (the Twin Creek advance) about 5000 yr B.P. At Linnet Lake, the onset of cold conditions begins at 5000 yr B.P. and terminates about 4000 yr B.P. (Christensen and Hills, this volume). Zone II at Linnet Lake is characterized by high percentages of arboreal pollen, notably *Abies*, and a decrease in grassland indicators. The *Abies* peak is interesting, because fir pollen probably travels short distances (Bujak 1974), and is found in the upper subalpine forest. This suggests that subalpine forest dropped below modern elevations between 5000 and 4000 yr B.P. in Waterton. In the middle of Zone II at Linnet Lake, there is a decrease in *Picea* and *Abies*, and an increase in *Pinus contorta*. *Pinus contorta* is a fire-successional species in the subalpine forest, and apparently, for a period of a few hundred years, fires were common in Waterton. Christensen and Hills (1971) have suggested that this long period of subclimax forest is the result of forest fires set deliberately by prehistoric people, who wished to increase the ungulate population of the area by destroying the encroaching, unproductive forests. Burning of the Plains by aboriginal groups is known from historical records (Arthur 1968), and such techniques could have been applied in the mountains.

Alley (1972) has documented a post-Altithermal cool period at Callum Bog, but deposits above Mazama Ash are relatively thin, and Neoglacial periods cannot be defined easily. Bujak (1974) has demonstrated the presence of Mazama tephra in Goat and Lost Lake (cirque lakes), Waterton Lakes National Park, indicating that although climates were cooler than previously, they were not sufficiently cool to produce cirque glaciation at these lakes.

Subzone 3b at Crowsnest demonstrates a return to warmer conditions, probably between 4000 and 3000 yr B.P. and correlates with Zone II at Linnet Lake (Christensen and Hills, this volume). In Zone II, *Picea* and *Abies* are reduced from the previous Zone III, and pollen derived from grasslands increases in a similar manner to the situation at Crowsnest.

A final piece of evidence for a warm period at this time in the southern Alberta Rockies is a fragment of wood recovered from a lateral moraine on the Peyto Glacier, Banff National Park. The wood, from a large coniferous tree, has been dated to 2880 ± 170 yr B.P. (Lowdon *et al.* 1971). The moraine is 200 m above modern upper timberline, showing that at 2900 yr B.P. forests existed higher in the mountains than today.

3000 yr B.P. to the Present

The change from Subzone 3b to Zone 4 in the Crowsnest core is dated at 2750 ± 410 yr B.P., and is marked by a sudden increase in coniferous pollen, and a reduction in the amount of grassland pollen. Zone 4 may contain two particularly cool and wet periods. The beginning of Zone 4 correlates with the beginning of Zone I at Linnet Lake, dated at 2830 ± 150 yr B.P. (Christensen and Hills, this volume). This zone shows an increase in *Picea* pollen, while the pollen of *Pinus* and grassland indicators decreases from Zone II. As at Crowsnest, encroachment of subalpine forest can be suggested. At Goat Lake, in the upper subalpine forest at Waterton, avalanche debris occurs in deposits dating to this period. The end of the avalanche phase has been dated to 1265 ± 150 yr B.P. (Bujak 1974). Data from the two lakes suggest that a period of increased precipitation occurred in Waterton from about 3000 to 1300 yr B.P. This probably correlates with the early part of Zone 4 at Crowsnest. Schweger *et al.* (1981) consider this interval began about 2800 yr B.P. in the Parkland area of central Alberta.

In the last few centuries, there was renewed glacial activity in the Rockies. Ice advances occurred in Banff and Jasper National Parks in the late nineteenth and early twentieth centuries. These advances were often quite extensive, and may have destroyed moraines left behind by less extensive advances at ca 5000 yr B.P. and *ca* 3000 yr B.P.

Summary: Palaeoecology of the Southern Alberta Rockies

The Holocene sequence in the Alberta Rockies generally, and in the Crowsnest Pass in particular, is by no means well established. The early postglacial sequence is only beginning to be revealed. The palynological sequence for the last 6600 years is moderately well known, but moraine sequences are not. In spite of these problems, a tentative sequence can be proposed.

Deglaciation of mountain valleys took place at different times in Alberta. Crowsnest Pass appears to have been ice-free very early, and was certainly supporting groups of animals during Classical Wisconsin time, if a date of 22,700 ± 1000 yr B.P. from a cave overlooking Crowsnest Lake is correct (Burns 1975). According to Alley's Callum Bog sequence, the vegetation in glacial times was a steppe type, with large areas of muskeg and small clumps of trees. The last Wisconsinan glaciation was the equivalent of Pinedale III in the United States Rockies, and dates 11,500 to 10,500 yr B.P. Following this, climate became warmer, and forests gradually colonized the mountain slopes, possibly reaching their modern elevations about 9000 yr B.P. A brief climatic reversal occurred about 8500 yr B.P., and for 500 years the climate was cooler and wetter. Lower treelines may have been slightly depressed at the time.

By 7500 yr B.P., glaciation was over, and the warming trend continued. From about 7500 to 5000 yr B.P. (the Altithermal), climate was warmer and drier than today. Grasslands in the mountain valleys expanded, as did the newly established montane forest zone. As a result of these factors, mountain areas may have been able to support larger populations of ungulates.

Following the Altithermal, a series of climatic fluctuations occurred. Cold periods, possibly accompanied by renewed glacial activity, reduced the elevation of vegetation zones during three periods: 5000 to 4000 yr B.P., 3000 to 1300 yr B.P., and the last few centuries. These periods are best seen in pollen diagrams which are sensitive to changes at the grassland/subalpine forest boundary (e.g. Linnet Lake, Crowsnest Lake). In lakes which are not so sensitive, palynology shows a trend to cooler conditions over the last 5000 years (Bujak 1974).

COMPARISON OF THE ALBERTA ROCKIES HOLOCENE SEQUENCE WITH OTHER AREAS

Comparisons of the Alberta sequence with other areas can be made in order to assess the extent of climatic events, and to see if similar events occurred in adjacent areas. In the following sections, major periods in the Alberta sequence are compared with the same periods in the Rockies as far south as Colorado, in southeastern British Columbia, in central Alberta, and in the Canadian Plains.

Early Postglacial to 8500 yr B.P.

A number of high-altitude studies in the Colorado Rockies show the development of vegetation in the changing postglacial environment. Benedict (1975) has defined a glacial sequence for the Colorado Front Range. The latest Pinedale stade is Santana Peak, with moraines dating between 11,000 and 10,000 yr B.P. Following this, there was an ice recession and forests advanced rapidly up mountain slopes, so that by 8000 to 9000 yr B.P.

the upper treeline was like that of today. Similar dates have been proposed for the latest Pinedale advance in Wyoming (Birkeland 1973; Currey 1973).

Andrews *et al.* (1975) have combined lithological and palynological evidence from the San Juan Mountains of Colorado. Deglaciation took place in the cirques before 10,000 yr B.P., and ice did not reappear. Thus cold periods in the Holocene in the San Juan Mountains were not accompanied by glaciation, and problems of later moraines obliterating earlier moraines do not occur. Palynological evidence suggests a warm period following 10,000 yr B.P., with upper treelines reaching modern limits at about 8500 yr B.P. Maher (1972) has documented a similar situation farther north in Colorado, where treelines gained altitude from 9500 to 8000 yr B.P.

In Yellowstone National Park, on the Montana/Wyoming border, two pollen sequences indicate late glacial and postglacial vegetation (Baker 1970; Waddington and Wright 1974). Baker's sequence shows a glacial vegetation similar to modern alpine tundra, although the sampling site is now within subalpine forest. From about 11,500 to 10,000 yr B.P., a *Pinus contorta/Pinus albicaulis* (whitebark pine) forest began to invade the tundra. This forest type is indicative of a cooler, drier climate than today's, but warmer than the situation at 11,500 yr B.P. At about 10,000 yr B.P., the forest loses much of the whitebark pine, suggesting increased warmth, as whitebark pine is today found at higher elevations than lodgepole pine.

Waddington and Wright (1974) have shown the existence of a steppe environment in the same area from 14,300 to 11,600 yr B.P. From 11,600 to approximately 9000 yr B.P., treelines rose in elevation, but the presence of whitebark pine throughout the period has been interpreted to show that treelines were not as high as today. A warming trend begins about 9000 yr B.P.

In the now arid area of southeastern Idaho, Bright (1966) has documented a late glacial coniferous forest, indicative of a colder and wetter period than today. This forest has been dated from 12,000 to 11,400 yr B.P., when a warming period began, but temperatures were still lower than today. By 10,300 yr B.P., a vegetation similar to that of today had appeared, and lasted until 8500 yr B.P.

Postglacial palynological sequences were obtained in northern Idaho and northwestern Montana before the invention of radiocarbon dating, by Hansen (1939, 1948). Early postglacial forests in both areas were composed of *Pinus albicaulis* and *Pinus contorta*.

Recent work (Waddington and Wright 1974) from western Montana has produced a similar early postglacial sequence to Yellowstone. The earliest zone at Lost Trail Pass Bog in the Bitterroot Mountains terminates at about 11,500 yr B.P., and shows a steppe vegetation in late-glacial time. From about 11,500 to 7000 yr B.P. an open forest occupied this area. As in other areas, *Pinus albicaulis* was an important constituent.

Early, undated sequences in central British Columbia show that a forest indicative of cool, wet conditions preceded the Altithermal (Hansen 1955).

From the arid Okanagan Valley in southeastern British Columbia, Alley (1975) has shown that a spruce and pine forest developed after a late deglaciation. By 8400 yr B.P., the forest had been replaced by a grassland with high amounts of *Artemisia*. This suggests that the area was very dry, as it is today.

The postglacial sequence east of the Rockies has been described for Lofty Lake in northern central Alberta (Lichti-Federovich 1970). The earliest zone is interpreted as a pioneering vegetation which briefly colonized the deglaciated landscape at about 11,500 yr B.P. Pollen suggests that the first plants were aspen, willow, herbs and grasses. This vegetation was followed rapidly by a spruce forest, which remained in the area until 9200 yr B.P., when a deciduous forest displaced the spruce. The forest, containing birch, aspen and hazel seems to have been an equivalent of the modern aspen parkland, which lies between the Plains grassland and boreal forest. This vegetation remained until 7500 yr B.P. The Lofty Lake sequence documents the northward migrations of various vegetation zones during the postglacial. Apparently, by 9200 yr B.P. a climate similar to that of today was present.

In Saskatchewan and Manitoba, a similar sequence has been observed in several pollen studies (Ritchie 1976). Following a brief interval of colonizing vegetation, spruce forests dominated many southern areas of the provinces until 10,500 yr B.P., when grasslands first appear on what is now the Plains.

By 10,500 yr B.P., in most places, vegetation zones were moving higher in elevation, or were drifting north in areas of low relief. In the early postglacial, forests were established in most mountainous areas. In some places, these forests were not composed of the same species as today. In particular, the presence of whitebark pine suggests somewhat cooler conditions. By 9000 yr B.P., the altitudinal or latitudinal position of vegetation zones was often equivalent to the modern situation.

A Cold Period from 8500 to 8000 yr B.P.?

As noted previously, there is equivocal evidence for a cold period in Alberta between 8500 and 8000 yr B.P. Although a climatic fluctuation can be observed at that time in Alberta, its effect upon vegetation is probably less drastic than previously thought.

The most convincing evidence for a cold, wet period then comes from Colorado, although Graf (1971) first dated a moraine to 8000 yr B.P. in the Beartooth Mountains of Wyoming and Montana. Benedict (1973) recorded a layer of sand in a peat bog situated in a cirque on the Colorado Front Range. He suggests that the change in lithology from peat to sand was the result of glacial outwash sediments entering the bog at about 8000 yr B.P. However, Maher's (1972) studies show no important changes of elevation of upper treeline in the Front Range at this time. Andrews *et al.* (1975) suggest that there was a minor lowering of upper treeline in the northern San Juan Mountains, Colorado, about 8300 yr B.P., based on bog stratigraphy and palynology. A similar fluctuation (dated from 8500 to 6700 yr B.P.) has been noted for the La Plata Mountains of Colorado by Peterson and Mehringer (1976). It should be noted that the last two studies were from sites near upper treeline, which would be expected to record most changes in elevation of that ecotone.

In southeastern Idaho, there is no evidence for a cool, wet period at this time, and Bright (1966) suggests that the Altithermal begins at 8400 yr B.P.

Nor is there evidence from Yellowstone or the Bitterroot Mountains for changes in treeline or forest composition then (Baker 1970; Waddington and Wright 1974; Mehringer *et al.* 1977).

In southeastern British Columbia, there is no palynological evidence to suggest the presence of such a cool, wet period, and Alley (1975) dates the beginning of the Altithermal to 8400 yr B.P. in the Okanagan Valley. Assuming that Hansen's palynological sequences extend back to 9000 yr B.P., evidently no changes in forest composition or elevation occurred before the onset of the Altithermal in northern Idaho, northwestern Montana, central British Columbia or central Alberta (Hansen 1939, 1948, 1949, 1955). No significant reversals of warming trends can be observed for this period at Lofty Lake (Lichti-Federovich 1970), nor can reversals be seen in Manitoba and Saskatchewan (Ritchie 1976).

In conclusion, evidence remains equivocal for the 8500 to 8000 yr B.P. period. If a climatic change did occur at this time, it can only be recorded at sensitive sites, such as those in Colorado, where sampling areas are close to major ecological boundaries. This suggests that any climatic alteration that may have occurred was quite minor, although all vegetation boundaries may have shifted slightly. Certainly, no major ecological change took place.

The Altithermal

There is considerable evidence for a warm period during mid-Holocene times. Petersen and Mehringer (1976) have shown that upper treeline gained altitude from 6500 to 5800 yr B.P. Andrews *et al.* (1975) extend the period of high upper treelines to cover the 7700 to 5000 yr B.P. interval. Benedict (1975) has suggested that the Altithermal in the Colorado Rockies can be divided into a dry (7500 to 6500 yr B.P.) and wet (6500 to 5000 yr B.P.) period. Treelines rose during the former, but lost altitude in the latter. Benedict's wet Altithermal is based upon very little evidence, and he may have overemphasized its presence to support his view of human abandonment of the Front Ranges during the warm Altithermal. (Benedict has suggested that the Front Ranges were not occupied from 7500 to 6500 yr B.P., because lack of spring and summer meltwater rendered the area too susceptible to drought.) However, reoccupation takes place after 6500 yr B.P., and it is therefore necessary for Benedict to postulate climatic change to account for this. Following Reeves (1973), we suggest that Altithermal abandonment is a dubious concept - a view which is being increasingly supported by evidence from the Northwestern Plains.

Studies from Wyoming, Montana and Idaho (Hansen 1939, 1948; Bright 1966; Baker 1970; Waddington and Wright 1974; Mehringer *et al.* 1977) show that forest species composition changed during a mid-Holocene warm period, which is widely dated by the presence of Mazama tephra. The period is poorly dated at Yellowstone, although it apparently terminates at about 5000 yr B.P. (Baker 1970). Mehringer *et al.* (1977) date the Altithermal between 7000 and 4800 yr B.P., whereas Bright (1966) suggests 8400 to 3100 yr B.P., a period when increased aridity can be seen in southeastern Idaho.

In central British Columbia, forest composition changes during a period around the time of Mazama tephra deposition. Warmer, drier conditions are indicated (Hansen 1955). In the Okanagan Valley, arid conditions last from 8400 to 6600 B.P. (Alley 1975).

Similar changes to those occurring in Crowsnest Pass can be seen in a core from Edson, in west-central Alberta (Hansen 1949). Just before Mazama tephra deposition, amounts of *Picea* pollen decrease, whereas *Pinus*, Gramineae and Chenopodiineae increase. This suggests that forests in the Rockies west of Edson were reduced in area, and that local Foothills forests were being replaced by areas of grassland.

The sequence from Lofty Lake (Lichti-Federovich 1970) seems to disagree with many other studies as to the duration of the Altithermal. Several studies date the end of the warm period to about 5000 yr B.P. Lichti-Federovich suggests a date of about 3500 yr B.P. for the end of Zone 4 at Lofty Lake. We suggest that Zone 4 can be subdivided into three periods. The first of these, the Altithermal, ends at about 5200 yr B.P. when an increase in *Betula* corresponding to a decrease in grassland species occurs. This would date the expansion of the grasslands north into the deciduous forest zone to between 7500 and 5200 yr B.P. It is between these two dates that grassland indicators such as Gramineae, Chenopodiineae, *Ambrosia*, and *Artemisia* reach high values. Following the Altithermal, we postulate a cool period followed by a shorter warming period, which ends at about 3500 yr B.P. Thus the originally defined Altithermal at Lofty Lake can be reinterpreted as two warm periods, separated by a short, cool, moist period.

Ritchie (1976) suggests that grasslands expanded beyond their modern boundaries north into the deciduous forest during the postglacial, and reached their maximum by about 6500 yr B.P. Following this, deciduous forest and boreal forest zones began to drift south into grassland areas. The present study suggests that grasslands may have continued to expand until about 5000 yr B.P. and that southward movement of the boreal forest occurred only after this date.

Neoglacial Stades and Interstades

Three cool, wet periods may characterize the Crowsnest sequence. The first may date from 5000 to 4000 yr B.P. The second begins at 3000 yr B.P. The third occurred during the last few centuries.

Not all pollen cores are sufficiently well dated to define precisely stades and interstades, and palynological sequences need not necessarily be expected to record such events, if their effect upon vegetation is minimal. Baker (1970) suggests that cool, moist periods occur near Yellowstone at 5000 yr B.P. and 2800 yr B.P. These dates are similar to those from Waterton and Crowsnest. In southeastern Idaho, Bright (1966) notes a cool period which dates from 3100 to 1700 yr B.P. - also similar to the southern Alberta dates for the second stade.

A cool, wet period was dated at 4800 to 4000 yr B.P. in the Bitterroot Mountains of Montana. The forest around Lost Trail Pass Bog lost its Douglas fir at this time, while pine and spruce increased (Mehringer *et al.* 1977).

In the Okanagan Valley, three stades can be defined, based upon increases in deciduous forest (Alley 1975) from 6000 to 4000 yr B.P., 3200 to 2000 yr B.P., and from 1500 yr B.P. to the present.

As suggested earlier, the middle part of the warm period defined at Lofty Lake may show decreasing temperatures and increased precipitation. If this interpretation is correct, the earliest post-Altithermal cool interval would be at about 5200 yr B.P. Following this, was a brief resurgence of grassland, possibly from 4000 to 3500 yr B.P. From 3500 yr B.P. to the present, modern deciduous forest was present at Lofty Lake (Lichti-Federovich 1970).

Southward movements of vegetation boundaries occurred in Saskatchewan and Manitoba from 6500 to 2500 yr B.P. Since 2500 yr B.P., no major changes in vegetation position have occurred (Ritchie 1976).

Regularities in the post-Altithermal climatic sequence can be observed, and correlated with changes observed at Waterton Lakes National Park and in the Crowsnest Pass. The date for the end of the Altithermal is often placed at about 5000 yr B.P. Dates for the end of the first cool interval are less common, but can be extrapolated to approximately 4000 yr B.P. The onset of cooling at about 3000 yr B.P. is fairly well dated in many areas, as are changes in vegetation for the same period. Following this, dates agree less frequently for stades and interstades, although in most areas a recent Neoglacial event is evident.

SUMMARY

In the preceding sections we show that the sequence of environmental events postulated for Crowsnest Pass can be equated with similar developments in other areas of Alberta and northwest North America. A generalized postglacial climate consists of an early warming trend, culminating in the Altithermal, with warmer temperatures and lower rainfall than today. Following this, two cold periods are readily documented, but the situation after 2000 yr B.P. is less certain.

Major changes affecting the prehistoric inhabitants of the Crowsnest Pass were probably few. After 10,500 yr B.P., grasslands were almost certainly established in Crowsnest Pass, and must have provided good grazing for ungulates. Forest zones gained elevation, perhaps reaching modern limits at about 9000 yr B.P. A brief decline in elevation may have occurred from 8500 to 8000 yr B.P., but the extent of the grasslands cannot have been seriously affected. Expansion of the grassland zone started shortly after this, and, by 7500 or 7000 yr B.P., the area had increased considerably. Then, subalpine forest moved up the valley sides, and increased herds occupied the grasslands and montane forest. Between 6000 and 5000 yr B.P., a change took place. Temperatures dropped, precipitation increased, and less productive subalpine forest expanded into the valley grasslands, reducing montane forest to a thin strip. At that time, vegetation zones may still have been somewhat higher in elevation than today. Following a second grassland expansion, cool, wet conditions reappeared, possibly reducing grasslands to an area approximately the same as today. With minor fluctuations in climate, this situation continued until today.

We emphasize that changes in the Crowsnest Pass have been largely quantitative. Undoubtedly the main factor in assessing the effects of environment on prehistoric man is the extent of the grasslands, because these provided habitat for the ungulates which were the subsistence base. Grasslands have not disappeared from the Crowsnest Pass at any time in the Holocene. Thus, there has always been a potential for a population of ungulates in the area - a potential probably unrivalled by any other valley in Alberta within the mountain front.

REFERENCES

Alley, N.F. 1972. The Quaternary history of part of the Rocky Mountains, Foothills, Plains, and Western Porcupine Hills, southwestern Alberta. Ph.D. thesis, University of Calgary, Calgary, Alberta. 201 pp.

______. 1975. Post Pleistocene climatic changes and their influence on the landscape, Okanagan Valley, British Columbia. Paper presented at the 1975 Conference of the Archaeological Association (Chacmool), University of Calgary, Calgary, Alberta.

______. 1976. The palynology and paleoclimatic significance of a dated core of Holocene peat, Okanagan Valley, southern British Columbia. Canadian Journal of Earth Sciences 13:1131-1144.

Andrews, J.T., Carrara, P.E., King, F.G., and R. Stuckenrath. 1975. Holocene environmental changes in the Alpine zone, Northern San Jan Mountains, Colorado: evidence from bog stratigraphy and palynology. Quaternary Research 5:173-197.

Arthur, G.W. 1968. Southern Montana. In: The Northwestern Plains: a symposium. Edited by: W.W. Caldwell. The Center for Indian Studies, Rocky Mountain College, Billings, Montana, Occasional Papers No. 1:51-62.

Baker, R.G. 1970. Pollen sequence from Late Quaternary sediments in Yellowstone Park. Science 168:1449-1450.

Benedict, J.B. 1973. Chronology of cirque glaciation, Colorado Front Range. Quaternary Research 3:585-589.

______. 1975. Prehistoric man and climate: the view from timberline. In: Quaternary Studies. Edited by: R.P. Suggate and M.M. Cresswell. Royal Society of New Zealand, Wellington. pp. 67-74.

Birkeland, P.W. 1973. Reinterpretation of the type Temple Lake moraine. Geological Society of America, Abstracts with Programs 5(6):465-466.

Bright, R.C. 1966. Pollen and seed stratigraphy of Swan Lake, S.E. Idaho. Tebiwa 9(2):1-47.

Bujak, C.A. 1974. Recent palynology of Goat Lake and Lost Lake, Waterton Lakes National Park. M.Sc. thesis, University of Calgary, Calgary, Alberta.

Burns, J.A. 1975. Eagle Cave and the search for early man in Alberta. Paper presented at the 1975 meeting of the Canadian Archaeological Association, Thunder Bay, Ontario.

Christensen, O.A., and L.V. Hills. 1971. Palynologic and paleoclimatic interpretation of recent sediments, Waterton Lakes National Park, Alberta. Report to National and Historic Parks Branch, Department of Indian Affairs and Northern Development, Ottawa. 15 pp.

______. (This volume). Palynologic and paleoclimatic interpretation of Recent sediments, Waterton Lakes National Park, Alberta.

Currey, R.R. 1975. Late glacial and Neoglacial chronology - a call for resoluation. Geological Society of America, Abstracts With Program 5(6):475.

Driver, J.C. 1978. Holocene man and environments in the Crowsnest Pass, Alberta. Ph.D. thesis, University of Calgary, Calgary, Alberta.

Fergusson, A.J., and L.V. Hills. (This volume). A palynological record, upper Elk Valley, British Columbia.

Fergusson, A.J. and G.A. Osborn. 1981. Minimum age of deglaciation of the upper Elk Valley, British Columbia. Canadian Journal of Earth Sciences 13:1635-1636.

Graf, W.L. 1971. Quantitative analysis of Pinedale landforms, Beartooth Mountains, Montana and Wyoming. Arctic and Alpine Research 3(3):253-262.

Hansen, H.P. 1939. Pollen analysis of a bog in northern Idaho. American Journal of Botany 26:225-228.

______. 1948. Postglacial forests of the Glacier National Park region. Ecology 29:146-152.

______. 1949. Postglacial forests in west central Alberta, Canada. Bulletin of the Torrey Botanical Club, 76:278-298.

______. 1955. Postglacial forests in south central and central British Columbia. American Journal of Science 253:640-658.

Harris, S.A., and J. Howell. 1977. Chateau Lake Louise moraines - evidence for a new Holocene glacial event in southwest Alberta. Bulletin of Canadian Petroleum Geology 25(3):441-466.

Harris, S.A., and R.R. Waters. 1977. Late Quaternary history of southwest Alberta: a progress report. Bulletin of Canadian Petroleum Geology 25(1):35-62.

Holloway, R.G., V.M. Bryant Jr., and S. Valastro. 1981. A 16,000 year pollen record from Lake Wabamun, Alberta. Palynology 5:195-207.

Jackson, L.E., Jr., G.M. MacDonald, and M.C. Wilson. 1982. Paraglacial origin for terraced river sediments in Bow Valley, Alberta. Canadian Journal of Earth Sciences 19(12):2219-2231.

Lichti-Federovich, S. 1970. The pollen stratigraphy of a dated section of Late Pleistocene lake sediment from central Alberta. Canadian Journal of Earth Sciences 7(3):938-945.

Lowden, J.A., I.M. Robertson, and W. Blake, Jr. 1971. Geological Survey of Canada radiocarbon dates XI. Radiocarbon 13(2):255-324.

Luckman, B.H., and G.D. Osborn. 1979. Holocene glacier fluctuations in the middle Canadian Rocky Mountains. Quaternary Research 11:52-77.

Luckman, B.H., G.D., Osborn, and R.H. King. 1978. Chateau Lake Louise moraines - evidence for a new Holocene glacial event in southwest Alberta: a discussion. Bulletin of Canadian Petroleum Geology 26:398-402.

MacDonald, G.M. 1982. Late Quaternary paleoenvironments of the Morley Flats and Kananaskis Valley of southwestern Alberta. Canadian Journal of Earth Sciences 18:23-55.

Maher, L.J. 1972. Absolute pollen diagram of Redrock Lake, Boulder County, Colorado. Quaternary Research 2:531-553.

Mehringer, P.J., S.F. Arno, and K.L. Petersen. 1977. Postglacial history of Lost Trail Pass Bog, Bitterroot Mountains, Montana. Arctic and Alpine Research 9(4):345-368.

Mott, R.J., and L.E. Jackson, Jr. 1982. An 18,000-year palynological record from the southern Alberta segment of the classical Wisconsin "Ice-Free" Corridor. Canadian Journal of Earth Sciences 19:504-513.

Osborn, G.D., and M. Duford. 1976. An early Holocene glacial advance in the Canadian Rockies. Geological Society of America, Rocky Mountain Section, Abstracts with Programs 8(5):616.

Petersen, K.L., and P.J. Mehringer. 1976. Postglacial timberline fluctuations, La Plata Mountains, southwestern Colorado. Arctic and Alpine Research 8(3):275-288.

Reeves, B.O.K. 1972A. The archaeology of Pass Creek Valley, Waterton Lakes National Park. Volumes 1 and 2. National Historic Sites Service Manuscript Report 61. National Historic Parks Branch, Department of Indian Affairs and Northern Development, Ottawa.

______. 1972B. An inventory of archaeological sites in Waterton Lakes National Park. National Historic Parks Branch, Department of Indian Affairs and Northern Development, Ottawa.

______. 1973. The concept of Altithermal cultural hiatus in Northern Plains prehistory. American Anthropologist 75(5):1221-1253.

______. 1974A Crowsnest Pass archaeological project 1972 salvage excavations and survey. Archaeological Survey of Canada Mercury Series No. 19.

______. 1974B. Crowsnest Pass archaeological project 1973 salvage excavations and survey. Archaeological Survey of Canada Mercury Series No. 24.

______. 1974C. Prehistoric archaeological research on the eastern slopes of the Canadian Rockies 1967 - 1971. Canadian Archaeological Association Bulletin 6:1-31.

______. 1975A. Archaeological investigations: Village of Bellevue sewer and water system 1974. Report prepared for the Village of Bellevue, Alberta. Manuscript report on File, Alberta Culture, Edmonton.

______. 1975B. Early Holocene (*ca.* 8000 to 5500 B.C.) prehistoric land/resource utilization patterns in Waterton Lakes National Park, Alberta. Arctic and Alpine Research. 7(3):237-248.

Reeves, B.O.K., and J. Dormaar. 1972. A partial Holocene pedological and archaeological record from the Southern Alberta Rocky Mountains. Arctic and Alpine Research 4(4):325-336.

Richmond, G.M. 1965. Glaciation of the Rocky Mountains. In: The Quaternary of the United States. Edited by: H.E. Wright and D.G. Frey. Princeton University Press. pp. 217-230.

Ritchie, J.C. 1976. The late-Quaternary vegetational history of the Western Interior of Canada. Canadian Journal of Botany 54(15):1793-1818.

Rutter, N. 1972. Geomorphology and multiple glaciation in the area of Banff, Alberta. Geological Survey of Canada Bulletin 206:1-31.

Schweger, C., T. Habgood, and M. Hickman. 1981. Late Glacial-Holocene climatic changes of Alberta: the record from late sediment studies. In: The Impact of Climatic Fluctuations on Alberta's Resources and Environments. Edited by: K.R. Leggat and J.T.Kotylak. Proceedings workshop and annual meeting, Alberta Climatological Association, February 1981. Atmospheric Environment Service, Western Region, Environment Canada, Edmonton, Alberta. Report No. WAES-1-81:47-59.

Shaw, J. 1972. Pleistocene chronology and geomorphology of the Rocky Mountains in south and central Alberta. In: Mountain Geomorphology. Edited by: H.O. Slaymaker and H.J. MacPherson. Tantalus Press, Vancouver. pp. 37-46.

Stalker, A. MacS. 1969. A probable Late Pinedale terminal moraine in the Castle River Valley, Alberta. Bulletin of the Geological Society of America 80:2115-2122.

Stalker, A. MacS., and J.E. Harrison. 1977. Quaternary glaciation of the Waterton - Castle River region of Alberta. In: Cordilleran Geology of Southern Alberta and Adjacent Areas. Edited by: M.S. Shawa. Bulletin of Canadian Petroleum Geology 25(4):882-906.

Waddington, J.C.B., and H.E. Wright, Jr. 1974. Late Quaternary vegetational changes on the east side of Yellowstone Park, Wyoming. Quaternary Research 4:175-184.

Wright, H.E., Jr. 1967. The use of surface samples in Quaternary pollen analysis. Review of Palaeobotany and Palynology 2:321-330.

PALEOBOTANICAL EVIDENCE FOR CLIMATIC CHANGE IN SOUTHERN BRITISH COLUMBIA DURING LATE-GLACIAL AND HOLOCENE TIME

Rolf W. Mathewes[1]

INTRODUCTION AND HISTORICAL PERSPECTIVE

Significant improvements have occurred in the late-Quaternary paleobotanical database of southern British Columbia during the last decade. There are now several well-dated pollen diagrams from coastal and interior localities, as well as supporting paleoecological data from plant macrofossils, molluscs, and sediment stratigraphy at a few sites. These data now permit a tentative reassessment of paleoenvironmental reconstructions for this area, although our knowledge of this diverse and complex region is still only in its infancy.

Paleobotanical studies provide direct information about past vegetation changes. The causes of such changes can and have been variously interpreted, depending on the interests and biases of the investigators. Climate, because of its acknowledged importance in influencing broad-scale vegetation patterns, has been a favourite explanation in accounting for past changes in distribution or abundance of plant species. Recent reviews, however, point out that our knowledge of the role of climatic parameters in controlling non-cultivated plant distributions is poor in most cases, and the possible roles of migration rates, competition, pathogens, and herbivores should not be overlooked in paleoecological studies (Davis 1978; Birks 1981). Nevertheless, the general usefulness of pollen and plant macrofossils in making paleoclimatic inferences is well established (Webb 1980).

Difficulties in interpretation were already apparent in pioneering pollen analytical studies in the Pacific Northwest. In his first palynological study in British Columbia, Hansen (1940) argued that an early postglacial cool, moist period was succeeded by an interval warmer and drier than present. Later, Hansen (1947) evaluated the relative roles of natural succession, soil conditions, fire, and climate in the Puget Lowland of Washington. He ascribed only a limited influence to climate in controlling forest changes, emphasizing instead the importance of regional fire history and local soil conditions. A

---

[1] Department of Biological Sciences, Simon Fraser University, Burnaby, British Columbia V5A 1S6

period of warm, dry climate during the Holocene was still postulated, however, followed by a more recent return to a moister and cooler climate. The importance of climate was down-played more recently, when Hansen (1974, p.4) stated that the "thermal maximum" is not reflected in the Puget Lowland, and that "... the 12,000 years of recorded forest succession suggests normal forest succession in deglaciated terrain, with little climatic fluctuation." In his analysis of three bogs on Vancouver Island (Hansen 1950, p. 276), he came to a similar conclusion, stating that "It seems doubtful that the marine climate has fluctuated materially since the early postglacial amelioration of the pro-glacial influence."

In more continental regions, however, such as eastern Washington and south-central British Columbia, Hansen has always recognized a mid-postglacial period of warm, dry climate (Hansen 1955, 1967).

When the concept of a postglacial "Hypsithermal interval" between about 9000 and 2600 yr B.P. was proposed by Deevey and Flint (1957), they suggested that the Cordillera of Alaska and British Columbia did not fit the pattern of Europe and eastern North America. The evidence of several glacial advances since 7000 yr B.P., during some of the warmest parts of hypsithermal time, could not be reconciled with the evidence from more intensively studied regions (Deevey and Flint 1957, p. 184).

Heusser (1960), in his extensive study of pollen chronologies along the northwest coast, nevertheless adopted the new hypsithermal concept and, used it as a time-stratigraphic unit with variable boundaries in zoning his pollen diagrams.

Several problems make the pioneering pollen studies in British Columbia and the rest of the Northwest difficult to interpret and compare. An absence or paucity of radiocarbon dates made correlations difficult, and an early overemphasis of coniferous tree pollen records by Hansen makes it impossible to assess the relative importance of deciduous trees and most non-arboreal taxa. Another major difficulty is the absence of Cupressaceae pollen curves in studies published prior to 1973. This is particularly important since red cedar (*Thuja plicata*) is presently a major forest component, along with western hemlock (*Tsuga heterophylla*), in much of coastal British Columbia and Washington State. Its absence in early pollen diagrams is as important a gap as the well-recognized absence of poplar pollen in many continental areas where poplars (*Populus*) form an important part of the tree cover. Fortunately, this gap is now being filled, and a "cedar-type" record is now emerging along the northwest coast (Hebda and Mathewes, unpublished).

In keeping with the theme of this meeting, the main aim of my paper is to evaluate the paleobotanical data in southern British Columbia, and see what climatic inferences might be drawn from it. Much of my discussion will focus on a "critical period" in the Holocene, namely the evidence for a period suspected as being warmer and perhaps drier than the present, whether it is called a Hypsithermal, Altithermal, Xerothermic, Climatic Optimum, or just warmer and drier. Although British Columbia is at the centre of my discussion, references to work in the adjacent United States and Alberta are frequently necessary. The names and locations of paleobotanical study sites are given in Figure 1 and Table 1.

Useful summaries which serve as a starting point for paleoclimatic discussions are Clague (1981) for all of British Columbia and Hebda (1982) for the grassland regions of the dry southern interior and the adjacent United States. Hills and Sangster (1980) compiled published data that also includes the Northwest. Heusser (1977) provides a good summary of climatic changes during the Quaternary in adjacent Washington State.

LATE-GLACIAL VEGETATION AND CLIMATE

Approximately 13,000-12,000 yr B.P.

Information on vegetation and climate during the early recessional phase of Vashon Stade ice in British Columbia is sparse. Deglaciation on the southwest coast was in progress by about 13,000 yr B.P. (Clague 1981), and the oldest published palynological data from British Columbia comes from basal clay at Marion Lake (Figure 1) and nearby glaciomarine sediments (Mathewes 1973). Lodgepole pine-type pollen dominates these early assemblages (Figure 2), dated at between 11,920 ± 245 yr B.P. (I - 6857) and 12,350 ± 190 yr B.P. (I - 5960) at Marion Lake (Mathewes and Heusser 1981), and at 12,690 ± 190 yr B.P. (I - 5959) in the marine sediment (Mathewes 1973). The high pine values (up to 91% of total pollen and spores) are typical of other early deglaciation sites in the Pacific Northwest, and suggest a forest, or parkland forest, dominated by lodgepole pine. Needle and cone macrofossils confirm that pine was locally present here, and also on southeastern Vancouver Island near Parksville, where Terasmae and Fyles (1959) investigated glaciomarine sediments also dated around 12,000 yr B.P.

Small amounts of true fir or balsam (*Abies*), spruce (*Picea*), and alder (*Alnus*) pollen are also recorded, along with fern spores and pollen of a variety of shrubs and herbs. Climatically, these assemblages are hard to characterize, since they have a

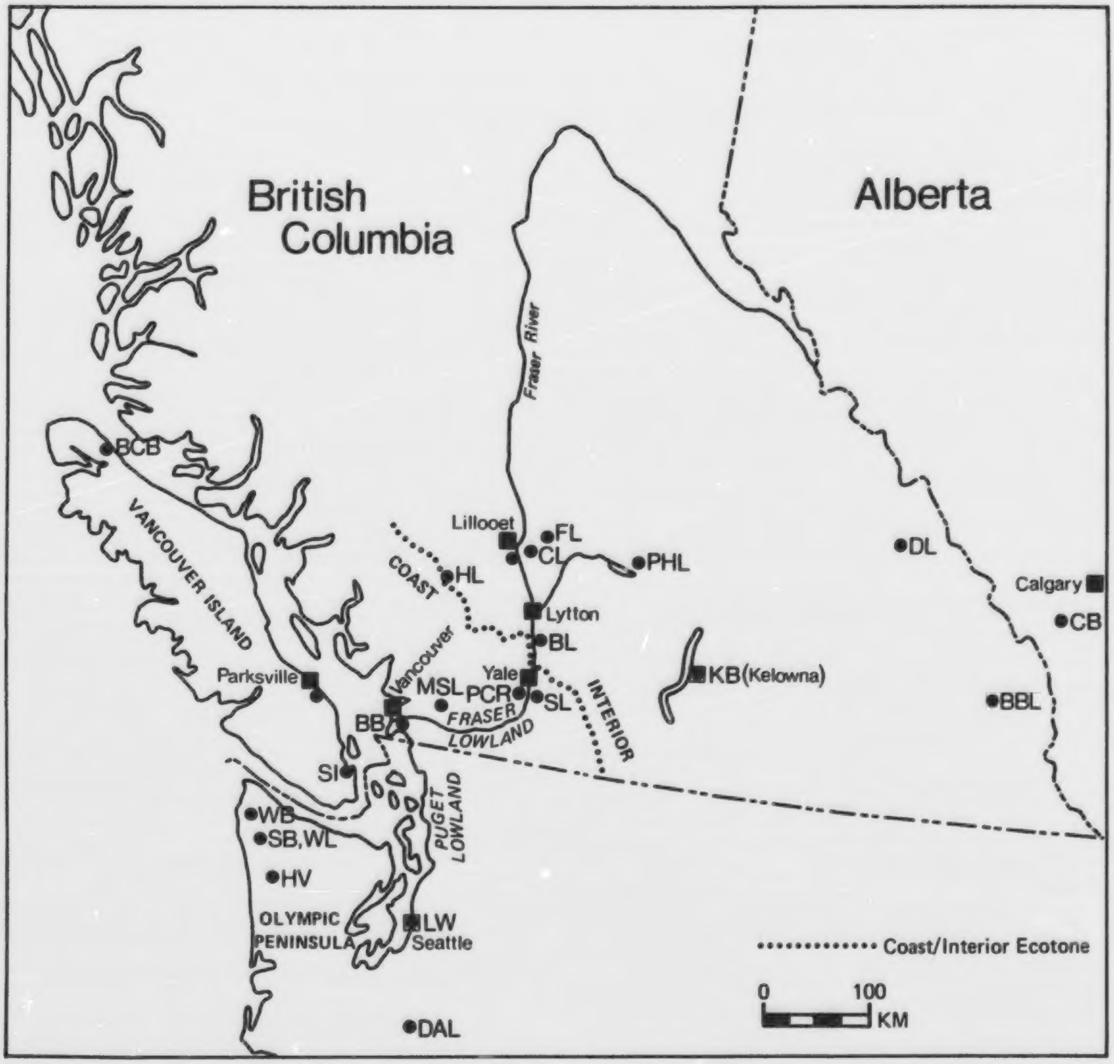

British Columbia
Alberta
Fraser River
BCB
VANCOUVER ISLAND
Lillooet
FL
CL
PHL
HL
COAST
Lytton
BL
DL
Calgary
CB
Parksville
Vancouver
MSL
PCR
Yale
SL
INTERIOR
KB (Kelowna)
BBL
BB
FRASER
LOWLAND
SI
PUGET LOWLAND
WB
SB,WL
HV
OLYMPIC PENINSULA
LW
Seattle
DAL
Coast/Interior Ecotone
0
100
KM

FIGURE 1: *Map of southern British Columbia and adjacent Washington, showing main sites and regions discussed in text. The approximate ecotone between coastal and interior ecosystems is shown for British Columbia.*

TABLE 1: *LIST OF PALEOBOTANICAL SITES DISCUSSED (SEE FIGURE 1), GIVING SITE NAME, AUTHOR, DATE, AND PUBLICATION STATUS.*

SITE NAME AND ABBREVIATION		AUTHOR	DATE	PUBLICATION STATUS
SOUTHERN BRITISH COLUMBIA				
BB	Burns Bog	Hebda	(1977)	Ph.D. thesis
BBL	Bluebird Lake	Hebda	(1982)	Details unpublished
BCB	Bear Cove Bog	Hebda		Unpublished
BL	Blue Lake	Mathewes		Unpublished
CL	Chilhil Lake	King	(1980)	M.Sc. thesis
DL	Dunbar Lake	Hazell	(1978)	M.Sc. thesis
FL	Finney Lake	Hebda	(1982)	Details unpublished
HL	Horseshoe Lake	Mathewes		Unpublished
KB	Kelowna Bog	Alley	(1976)	Published
MSL	Marion and Surprise Lakes	Mathewes	(1973)	Published
PCR	Pinecrest Lake	Mathewes & Rouse	(1975)	Published
PHL	Phair Lake	King	(1980)	M.Sc. thesis
SI	Saanich Inlet	Heusser	(1983)	Published
SL	Squeah Lake	Mathewes & Rouse	(1975)	Published
WASHINGTON				
DAL	Davis Lake	Barnosky	(1981)	Published
HV	Hoh Valley	Heusser	(1974)	Published
LW	Lake Washington	Davis	(1973)	Published
		Hansen	(1974)	Published
		Leopold *et al.*	(1982)	Published
SB	Soleduck Bog	Heusser	(1973)	Published
WB	Wessler Bog	Heusser	(1973)	Published
WL	Wentworth Lake	Heusser	(1973)	Published
ALBERTA				
CB	Chalmers Bog	Mott & Jackson	(1982)	Published

"pioneer" character about them that is difficult to interpret. Hansen (1947) has discussed the role of lodgepole pine as a pioneer species, and suggested that it is not a good climatic indicator.

Mathewes (1973, p. 2096) made the suggestion that these early environments may have been cool and continental, rather than humid coastal in climatic character, based on the presence of abundant *Shepherdia canadensis* pollen along with lodgepole pine, and possibly mountain alder (*Alnus incana*). *Shepherdia* has a spotty distribution in the driest parts of the south coast at present, whereas it is widespread and common in the interior of the province. Similarly, the occurrence of *Dryas drummondi* macrofossils in a late-glacial marine delta on southern Vancouver Island (Terasmae and Fyles 1959) also hints at a more continental climate at this time. Alternatively, since both *Dryas* and *Shepherdia*, as well as *Alnus*, are nitrogen-fixers, their early presence in areas now outside their normal ranges may simply reflect a pioneering situation followed by competitive displacements. A tentative designation of a "cool and dry climate?" is given in Figure 2, although further sites must be investigated, and other types of supporting data found for comparison. On northern Vancouver Island, at Bear Cove Bog, Hebda (in press) suggests from a bog section that climate may have been slightly cooler than today. On the Olympic Peninsula, July temperature appears to have been 1-1.5 °C cooler than today (Heusser 1973).

Compared to the coast, deglaciation in the interior of British Columbia appears to have begun rather late in most areas. Although conceding that parts of the southern interior may have been ice-free prior to 12,000 yr B.P., Clague (1981, p. 17) gives the oldest reliable postglacial interior date as 11,000 ± 180 yr B.P. (GSC - 909). Surprising, therefore, is a recently-cited sediment date of 13,000 yr B.P. (Heba 1982) from Finney Lake east of Lillooet (Figure 1). Hebda interprets the upland vegetation between 13,000 and 12,000 yr B.P. at this site as pioneering grassland, with sedges, grasses, and sage (*Artemisia*) growing under a climate that was as cool and dry as today (Hebda 1982, pp. 171,175). Further evaluation of this interesting site must await publication of the primary data.

Approximately 12,000 - 10,500 yr B.P.

There is general agreement that during this period on the coast, cool and relatively moist conditions prevailed (C.J. Heusser 1973, 1977; L.E. Heusser 1983; Hansen and Easterbrook 1974; Mathewes 1973; Mathewes and Rouse 1975; Mathewes and Heusser 1981).

Geological evidence suggests that a short-lived glacial resurgence (Sumas Stade) occurred in the Fraser Lowland around 11,300 years ago. Whether this event was climatically controlled or not is unclear, since non-climatic causes such as isostatic uplift and marine regression, resulting in grounding of glaciers, may be responsible (Clague 1981). The best paleobotanical evidence for a colder climate during Sumas time is Heusser's (1973, 1977) contention, from the Olympic Peninsula, that subalpine conditions prevailed between about 11,000 and 10,000 years ago. A peak in mountain hemlock (*Tsuga mertensiana*) pollen at this time is the main basis for this interpretation. Although peaks in mountain hemlock were also recorded around 10,500 - 10,000 yr B.P. at Marion and Surprise Lakes (Mathewes 1973), the values were considered to be too low (2 - 4%) to suggest a major climatic event. Heusser (1978) concluded from surface sample data that 1 - 2% mountain hemlock pollen in the Puget Lowland exemplifies carriage by wind of at least 50 km from the subalpine forests of the Coast Range. The peak value at Marion Lake (3.8%) also falls too late in time to correlate with a Sumas ice advance before 11,000 years ago. Although the evidence for a "glacial climate" is not clear from paleobotanical data, wetter and cooler conditions are suggested by the combined high values of mountain hemlock, balsam, spruce, and the small peak in western hemlock at the late-glacial/Holocene boundary.

Transfer functions were recently applied to the Marion Lake pollen data in order to estimate mean July temperature and annual precipitation trends over the past 12,000 years (Mathewes and Heusser 1981). The results are summarized in Figure 2. They support the previous qualitative interpretations of moist conditions cooler than the present. Similar data are available for sites in Washington (Heusser *et al.* 1980; Mathewes and Heusser 1981). Here too, the climatic reconstruction does not show a major cooling correlative with the Sumas event, although a small precipitation peak bounds the period between maximum mountain hemlock values and the initial small peak of western hemlock around 10,000 years ago.

Little has been published for this period in the interior of the province. Hebda (1982) estimates that lower valley slopes at Finney Lake and "Pemberton Hill" near Kamloops (Figure 1) may have supported closed stands of poplar or parkland with grasses and *Artemisia*, and that pine forests likely occupied higher elevations. He suggests a cool, moist climate (Hebda 1982, p.175), thus corresponding with the situation on the coast.

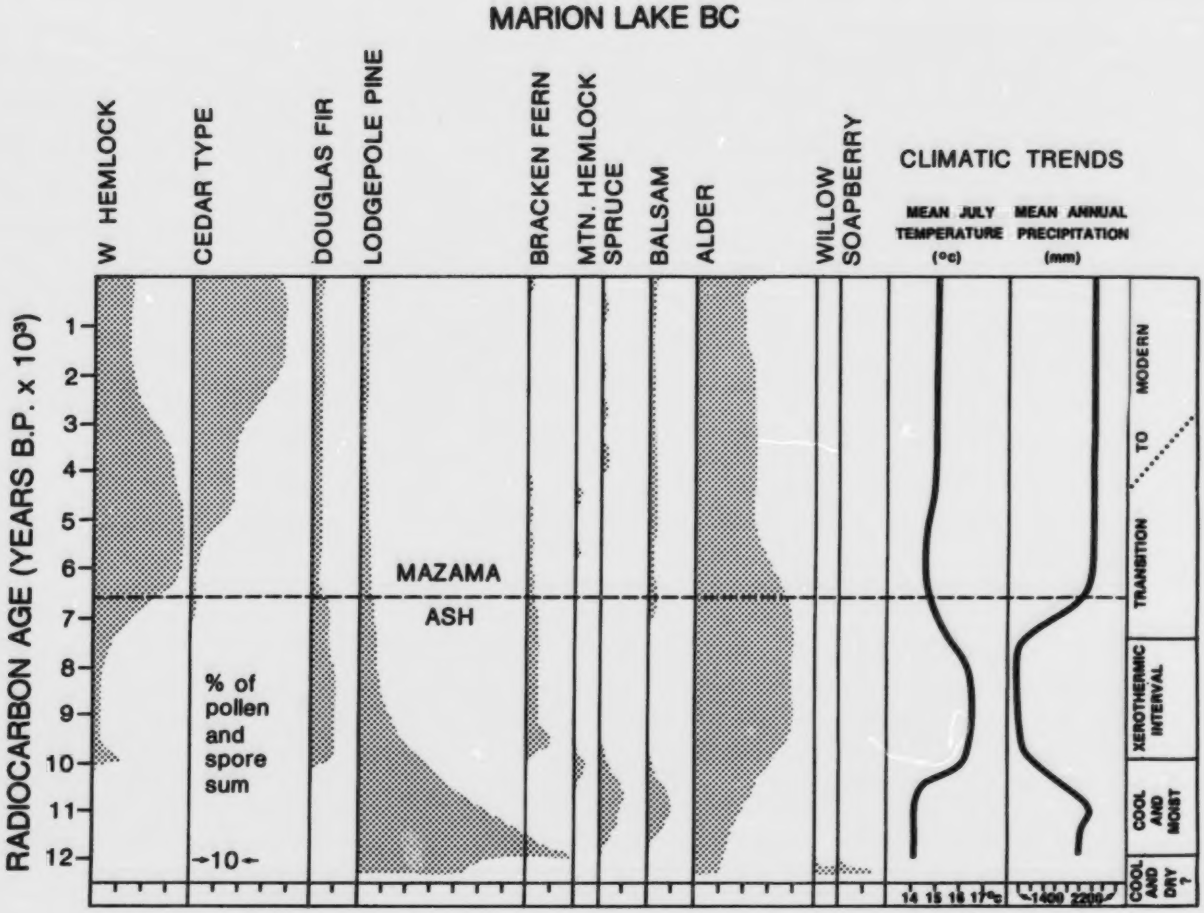


FIGURE 2: *Simplified pollen diagram and inferred climate history from Marion Lake, British Columbia. Curves are smoothed from Mathewes (1973) to show only main trends. Climatic trend-lines are simplified from Mathewes and Heusser (1981). The temperature and precipitation scales are not absolute - they only show strengths of trend changes.*

EARLY TO MID-HOLOCENE CLIMATE

Geological and botanical evidence for early to mid-Holocene climate in the Cordillera is complex and often contradictory. Radiocarbon dates on wood above present timberline in southern British Columbia attest to a period of warmer climate prior to 5200 years ago (Glague 1981). Some geological evidence, however, points to early Holocene glacial advances during this time, especially in the southern Rocky Mountains (Clague 1981). Pollen evidence in southwestern British Columbia and Washington State generally supports the existence of a warmer and/or drier period at some time during the Holocene, although the proposed dating varies considerably from locality to locality (Mathewes 1973; Mathewes and Heusser 1981; Hebda 1982). This is not unexpected in a region of extreme physiographic and climatic variability, especially in the light of much evidence that the Hypsithermal is a time-transgressive event in other parts of North America (Wright 1976).

Coast and Coast/Interior Transition

Between 10,500 yr B.P. and 10,000 yr B.P., a dramatic change is apparent in many pollen diagrams from south coastal British Columbia and Washington (Heusser 1960). In British Columbia, Mathewes (1973) radiocarbon-dated a major pollen zone boundary at Marion and Surprise Lakes at 10,370 years ago, and Mathewes and Rouse (1975) interpolated a similar age for lakes in the Fraser Canyon area. At Marion Lake (Figure 2), significant declines occur in lodgepole pine, spruce, balsam, and mountain hemlock during this interval. Douglas-fir (*Pseudotsuga menziesii*) appears abruptly, reaching its maximum Holocene values shortly thereafter. Western hemlock appears in low but significant quantities for the first time, and alder as well as bracken (*Pteridium aquilinum*) and other fern spores begin to increase. The suggestion that climatic warming may have favoured the expansion of Douglas-fir at this time was made by Mathewes (1973, p. 2099), although clear evidence of a subsequent period warmer and drier that the present was not recognized. In coastal Washington, Heusser (1977, p. 291) has argued that peak values of *Pseudotsuga* for extended periods in the past resulted from climatic conditions warmer and drier than the present.

A quantitive analysis of climatic trends at Marion Lake (Mathewes and Heusser 1981) suggests a rapid rise in summer temperatures, coupled with declining precipitation between 10,500 and 10,000 yr B.P. (Figure 2). The smoothed trend curves in Figure 2 clearly define a period of maximum July temperature and minimum precipitation between about 10,000 and about 7,500 years ago. This informally-named "early Holocene xerothermic interval" (Mathewes and Heusser 1981) is interesting because it supports the qualitative identification of warmer and drier early Holocene conditions at Pincrest and Squeah lakes in the Yale area (Figure 1) near the coast-interior transition (Mathewes and Rouse 1975), as well as in other coastal sites (Hansen and Easterbrook 1974; Heusser *et al.* 1980; Hebda, in press). Two unpublished lake sites recently studied by the author within the coast-interior ecotone (Blue and Horseshoe Lakes, Figure 1) exhibit a pattern similar to the Yale area, although evidence for xerothermic conditions persists until about 6,000 years ago. At Bear Cove Bog, near the present northern limit of Douglas-fir on Vancouver Island, the sudden appearance of significant quantities of Douglas-fir pollen, accompanied by bracken fern, indicates maximum dry and warm conditions between 8,800 and about 7,000 years ago.

A simplified composite pollen diagram for the Yale area shows the main qualitative evidence for an early xerothermic interval (Figure 3). The steep, rugged topography of the lower Fraser River Canyon (Figure 4) and the "sub-continental" climatic conditions here contrast with the wetter "maritime" conditions of the Marion Lake area to the west (Courtin *et al.* 1981). This climatic difference was likely present in the early Holocene, when high values of grass (Figure 5A) and *Artemisia* (Figure 5B) pollen indicate xeric, open areas within a forest of Douglas-fir and pine. Sunny rock outcrops with thin soil were likely the source of spikemoss spores (Figure 5C), since *Selaginella wallacei* commonly forms loose mats in such situations today. Such a rocky site is illustrated in Figure 6, from a very dry locality along Okanagan Lake near Kelowna. A surface sample of 20 amalgamated moss polsters in this locality exhibits grass and Douglas-fir pollen frequencies (ca. 5 - 6%) similar to those in the coastal early Holocene, although the high pine counts, absence of bracken, and very low alder values indicate that this site is not a modern analogue. However, *Selaginella* values are high (13%), and together with the grass values, suggest that physiognomically similar open patches might be interpreted during the early Holocene in the Yale area. Even at Marion and Surprise Lakes, where *Selaginella* spores are scarce in the cores, virtually all occurrences are from pre-Mazama sediment (Mathewes 1973), suggesting well-lit rock outcrops nearby.

The presence of abundant bracken spores (Figure 5D), together with high alder (Figure 5E) and Douglas-fir pollen (Figure 5F) are interpreted as representing a period of open forest conditions (Heusser 1973, 1974; Hebda, in press).

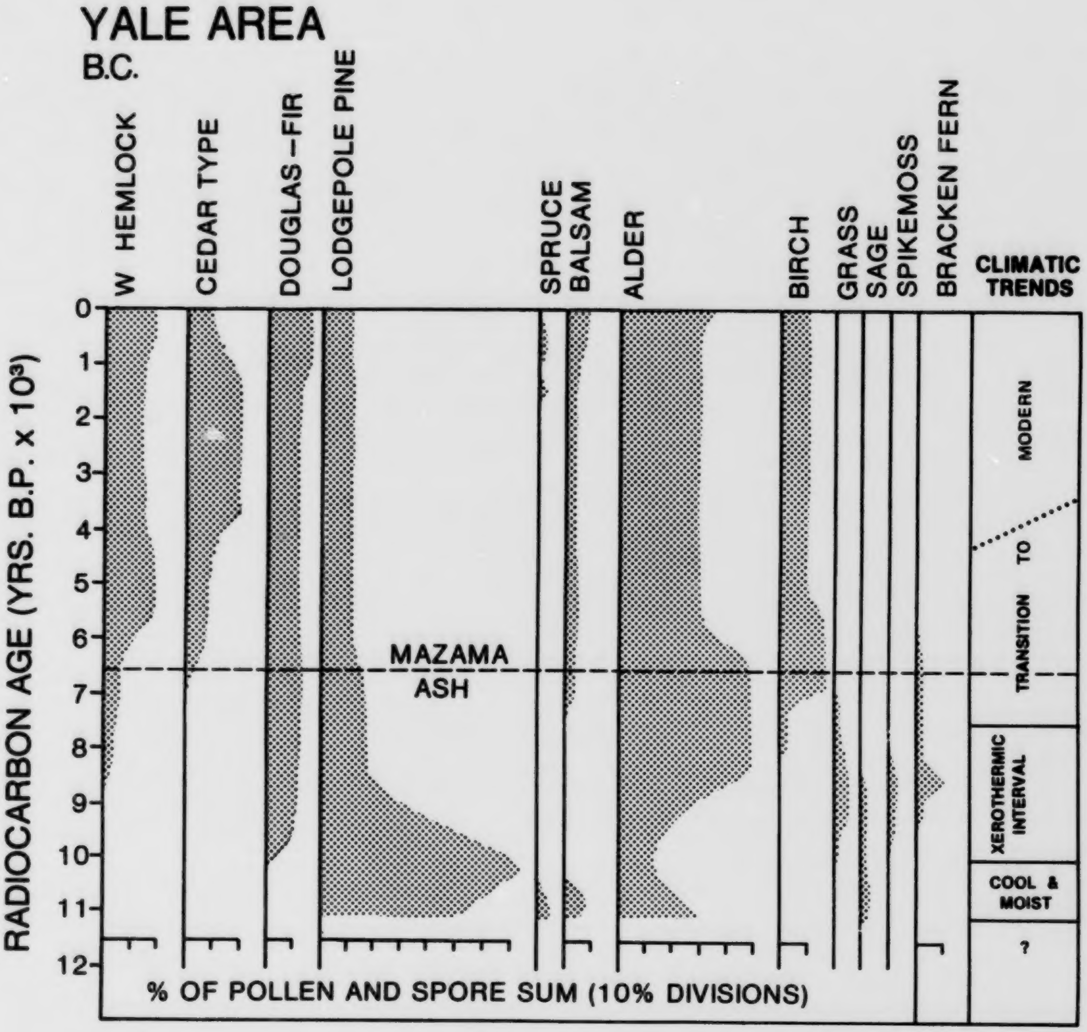


FIGURE 3: *Composite pollen diagram for Pinecrest and Squeah Lakes near Yale, British Columbia (Mathewes and Rouse 1975). Selected curves have been smoothed to show main patterns. Inferred climatic trends are summarized at right.*

FIGURE 4: *The Lower Fraser River Canyon near Pinecrest and Squeah Lakes. Looking east over the Fraser River to the northern Cascades. Although the area is in the coastal Western Hemlock zone, the well-drained bedrock slopes support an abundance of Douglas-fir (foreground).*

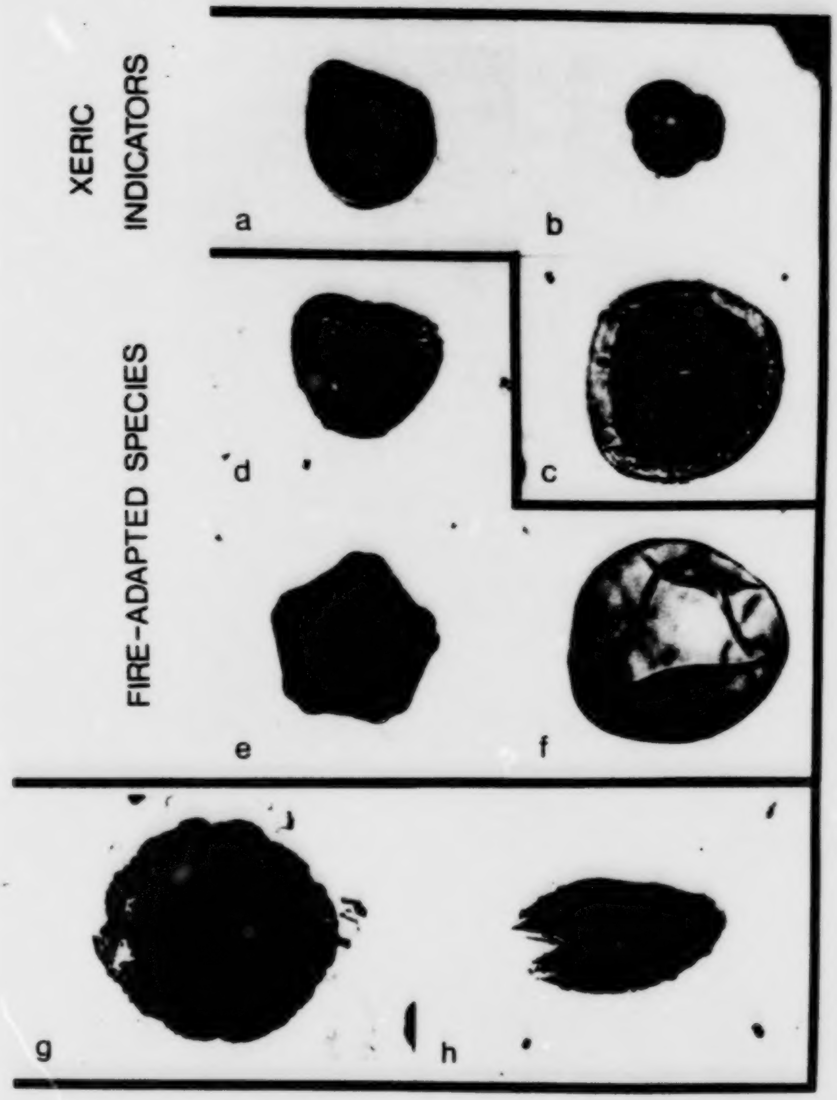


FIGURE 5: *Fossil pollen and spores from lake sediments in southwestern British Columbia. The three ecological groupings shown are not absolute, but serve as an interpretive guide for paleoclimatic assessment (see text).*

Magnifications are 500x for (f) and (g); 1000x for all others.
*(a) Grass pollen grain. (b) Sage (*Artemisia*) pollen in polar view. (c) Spikemoss (*Selaginella *cf.* wallacei*) microspore. (d) Bracken (*Pteridium aquilinum*) spore. (e) Alder (*Alnus*) pollen grain. (f) Douglas-fir (*Pseudotsuga*) pollen grain. (g) Western hemlock (*Tsuga heterophylla*) pollen in polar view. (h) Cedar-type (probably* Thuja plicata*) pollen, in the usual split and folded condition.*

Western hemlock and cedar type pollen are typically most abundant in the mid- to late Holocene on the south coast, and typify moist coniferous forest conditions.

FIGURE 6: *Bedrock slopes on the east shore of Okanagan Lake, near Kelowna, showing scattered ponderosa pine, Douglas-fir and shrubs in predominantly grassy vegetation with abundant mosses and* Selaginella wallacei.

BRACKEN, FIRE AND CLIMATE

Many palynologists have commented on the probable importance of past fires as determinants of vegetation in the Pacific Northwest, although solid evidence in the form of well-dated, sedimentary charcoal profiles is not yet published for the region. Most direct evidence of Holocene fires relates to reports of occasional sedimentary charcoal horizons (Mathewes 1973; Heusser 1973; Mathewes and Rouse 1975), although Leopold *et al.* (1982) cite a personal communication by Goldberg suggesting that sedimentary charcoal is present in Lake Washington between 10,000 and 7,000 years ago. The same zone is also characterized here by maximum bracken (32%), alder (62%), and Douglas-fir (20%) values. Alder, which is greatly overrepresented in pollen spectra, does appear to have been a more abundant component of early Holocene forests than it is at present. Changes in sedimentary lignin oxidation products in Lake Washington indicate a major angiosperm wood component that correlates well with the peak alder pollen frequencies (Leopold *et al.* 1982). At Soleduck Bog, Olympic Peninsula, Heusser (1973) points out that charcoal found in the early Holocene coincides with the period of maximum bracken abundance.

This early to mid-Holocene pattern of bracken abundance is an interesting phenomenon, inasmuch as late Holocene sediments do not exhibit such continuously high values. However, historical logging, slash-burning, and other forest-clearance activities have produced very significant recent increases in bracken spores at Lake Washington (Davis 1973) and Marion Lake (McLennan 1981), similar to those in the early Holocene. Early Holocene peaks in bracken are not an artifact of percentage pollen diagrams, since influx studies confirm the high relative values (Heusser 1974; Barnosky 1981; Mathewes, unpublished).

In the light of the palynologicial evidence, I suggest that there is an important connection between the presence of past communities with high bracken, alder, and Douglas-fir, and increased fire frequency. Fire frequency and climate are obviously also interrelated. The post-fire successional roles of alder and Douglas-fir are well known, and the following is a brief attempt to make bracken the third member of this fire-adapted triad (Figure 5).

Watt (1976) has summarized the ecology of bracken in Europe. It is a frost-sensitive plant, and its distribution in Britain is at least partly set by this characteristic. Its appearance during deglaciation in the Pacific Northwest may thus reflect a general amelioriation of climates in keeping with other evidence. It is also sensitive to lack of oxygen, a characteristic used to explain the absence of bracken in bogs, marsh, clay, and waterlogged soils in Europe. At the same time, it responds favourably to an adequate supply of water: on shallow and very dry soils its fronds are few and short.

Once established, a bracken colony proliferates vegetatively. Spore production is poorly understood, varying greatly with locality, and many colonies produce few if any fertile fronds (Conway 1957). Bracken shows a gradual decrease in fertility with increasing shade (Page 1976), supporting Conway's (1957) suggestion that spore production is probably lower in woodland areas than in the open. Spore production in fertile bracken is immense, with a single frond producing up to 3×10^8 spores (Conway 1957). Spores are released in peak numbers from late August through September, and reach their greatest aerial concentrations in warm, dry weather. It has been suggested that the highest deposition rates of spores would probably occur during the first rainfall following a warm, dry spell (Page 1976). At Marion and Surprise Lakes, however, McLennan (1981) found no significant differences in spores collected in roofed and unroofed traps, indicating that atmospheric rainout is probably not necessary to cause most deposition.

Significantly, establishment of bracken sporelings in Britain, France and Finland has mostly been recorded from burned sites (Page 1976). The reasons for this are poorly known, but microhabits created by burning probably aid colonization by spores. In south-coastal British Columbia, McMinn (1951) and others have pointed out the importance of bracken as a successional invader following fires.

These ecological features, together with the striking correspondence of Douglas-fir, alder, and bracken in pollen diagrams, suggest that fires were likely an important source of forest instability during the early to mid-Holocene. The combined influence of higher summer temperature, lower precipitation, and perhaps also higher lightning frequency would be expected to result in greater fire frequency in this region, even today characterized by a summer-dry climate.

A final comment about fern-spore abundance. The distinct pre-Mazama peak in monolete fern spores at Marion Lake (Mathewes 1973) is best interpreted as resulting from increased soil erosion and fluvial input. Recent studies confirm the importance of disturbances, whether by logging and fire, or by land clearance for urbanization and agriculture, in increasing the transport and depostion of soil-borne fern spores (McLennan 1981; Mathewes and D'Auria 1982).

REVERTENCE AND EARLY HOLOCENE CLIMATIC CHANGE

The relative roles of postglacial tree migration and climatic changes are a controversial topic in eastern North America (Davis 1976; Birks 1981). It is often assumed that tree distributions are in equilibrium with climate, but this assumption may not hold true, especially in early postglacial time when range extension is just beginning for many species. Thus, as Birks (1981, p. 127) points out, climatic inferences can only be made in equilibrium situations "...where species have migrated into an area and have subsequently retreated." "Retrogressions" or "revertences" in the distribution or abundance of individual taxa can therefore help in defining periods of climatic change.

Inspection of pollen diagrams suggest that such revertences may have occurred - particularly in western hemlock (Figure 5G) and possibly also in red cedar and balsam, as well as a few other taxa.

The general pattern in western hemlock is best seen in diagrams from the northwestern Olympic Peninsula (Heusser 1973) where high percentages (up to about 12%) prevail between 11,000 and 10,000 years ago at Wentworth Lake, then decline to values less than 2.1% before slowly increasing again around the mid-Holocene. Wessler Bog shows a similar sequence. The recently published diagrams of Heusser (1977), Barnosky (1981) and Leopold *et al.* (1982) also hint at this pattern. In British Columbia, a brief hemlock maximum exceeding 10% occurs in Marion Lake at 10,000 yr B.P., followed by values below 5% until around 8,000 years ago, where a slow increase begins (Figure 2). Influx studies are needed to confirm the reality of the pollen decline just after 10,000 yr B.P., which is highly suggestive of climatic change. The clearest example of hemlock revertence from British Columbia is provided by Hebda (in press), who showed a dramatic decline from around 40% to 11% of arboreal pollen at Bear Cove Bog in the early Holocene. A climatic cause is likely, since Douglas-fir and bracken peak during the hemlock minimum.

Similar patterns can be seen for the balsam curves at Marion Lake (Figure 2) and Surprise Lake (Mathewes 1973). A potentially climatically-significant revertence is suggested by the presence of *Thuja* seeds and foliage around 10,000 yr B.P. at Marion Lake (Mathewes 1973), when cedar-type pollen is virtually absent. Cupressaceae pollen (Figure 5H) continues to be rare until the Mazama ashfall, above which both pollen and macrofossils are well represented. Macrofossils in this study were only extracted from "detritus zones", where they were particularly abundant. A new core is currently being studied by N. Wainman in my laboratory to provide a continuous macrofossil record for comparison with the pollen evidence.

The modern ecology of western hemlock and red cedar is also instructive when compared to their Holocene histories. Both species have low drought tolerances, have a high "moisture optimum", and possess low fire resistance (Minore 1979). The characteristics of Douglas-fir are the opposite in each case, again suggesting a warmer, drier, climate with the possibility of increased fire frequency in the early Holocene.

Southern Interior

Hansen (1955) identified a warm dry period in south-central British Columbia between 7500 and 3500 years ago, with a "thermal maximum" around 6600 years ago. In his pollen analysis of Kelowna Bog (Figure 1), Alley (1976) found evidence of a warm dry interval, but between about 8400 to 6600 years ago. The Mazama volcanic tephra at 6600 yr B.P. correlates well with the return to moister and cooler conditions here, as it does on the coast. Alley's geomorphic studies corroborate his palynological data, with evidence for wind erosion and dune formation in the Okanagan Valley during the early Holocene, followed by dune stabilization.

In a detailed study of two Holocene lake cores near Lillooet, King (1980) analyzed pollen, plant macrofossils, and aquatic mollusc remains. At Chilhil Lake (Figure 1), the best evidence for early Holocene xerothermic conditions comes from pollen and seeds of "mudflat" plants, such as Chenopodiaceae and *Rumex*, which suggest that lake levels were very low between about 8000 yr B.P. and the Mazama ashfall. An abudance of pulmonate snails and seeds of shallow water aquatic plants also support the contention of low lake levels at this time. Phair Lake (Figure 1) is a small pond that began accumulating sediments just below the Mazama ash layer. A fibrous shallow-water peat with abundant Cyperaceae achenes and pollen of cattail (*Typha*) gives way to deeper-water marls around 5650 years ago. In the absence of geological data to the contrary, this sequence could be interpreted as representing increased precipitation around 7000 yr B.P., initiating a marsh phase at Phair Lake, followed by another precipitation increase leading to permanent pond formation after the Mazama ashfall. Supporting this climatic interpretation is the significant reduction of mudflat indicators at Chilhil Lake by 6100 yr B.P., signalling increasing lake levels. More studies of sediments and macrofossils are needed in the Pacific Northwest to identify and confirm lake level changes, which are useful indicators of climatic change (Wright 1966).

Hebda (1982) has reviewed the evidence for climatic changes relating to grassland distribution in the southern interior. He emphasizes the indicator value of non-arboreal pollen, especially pollen frequency curves of *Artemisia* and grasses in his recontructions, based on various sources and his own unpublished data from Finney, "Pemberton Hill" and Bluebird lakes (Figure 1). Hebda interprets the period between 10,000 and 8000 yr B.P. as a grassland maximum, with climate warmer and drier than present. This is followed by a "mesic grasslands period" between 8000 - 4500 years ago, reflecting a general trend to moister conditions and forest expansion during this time, although still warmer than the present. In areas of the adjacent intermontane United States, the pattern of vegetation succession and inferred climate is complex, suggesting a range of conditions from "similar to modern", to "warmer and drier", "warmer", or "moister" for time between 7500 to 4500 years ago (Hebda 1982, Table 3).

External correlations are also illuminating. White and Mathewes (1982) discuss the paleoclimatic implications of other studies in northwestern Canada. A number of basal lake and pond dates from this region suggest climatic change about 7,000 ± 500 yr B.P., beginning a wettening trend that culminates in the formation of permanent ponds and lakes on the Alberta Plateau between 5500 and 3000 years ago (White 1983; White and Mathewes 1982). The time-transagressive nature of suggested climatic shifts in the mid-Holocene is clear, and in view of the often contradictory evidence, I prefer to view this period as one of transition, both vegetatively and climatically, in the interior as well as on the coast (Figures 2, 3). Further studies of well-dated pollen and macrofossil sequences are necessary to clarify these patterns.

LATE HOLOCENE CLIMATE

Coast

Although the nature of early to mid-Holocene climate is controversial, there is widespread agreement among geologists and paleobotanists that the climate of the late Holocene became colder and wetter (Clague 1981). The beginning of this cooling trend, however, appears to be time-transagressive in the Cordillera, as discussed earlier. Geological evidence of neoglacial alpine glacier re-advances in the Cordillera is scattered throughout approximately the last 5000 radiocarbon years, with a particularly well-documented period of glacier expansion at 2300 to 3100 years ago (Clague 1981).

Paleobotanical indications of increasing wetness on the coast begin earlier than 5000 yr B.P. at some sites. Increasing pollen frequencies of western hemlock and cedar type (probably *Thuja*) around the time of Mazama ashfall (6600 yr B.P.) have been reported by Mathewes (1973), Mathewes and Rouse (1975) and Heusser (1983), for southwestern British Columbia (Figures 2, 3), and by Hansen and Easterbrook (1974) and Leopold *et al.* (1982) for northern Washington. Western hemlock and Sitka spruce (*Picea sitchensis*) increases were used by Hebda (in press) to infer cooling and increasing moisture after 7000 yr B.P. at Bear Cove, followed by the advent of modern hemlock-red cedar forest around 4000 - 3000 years ago. Based on the modern ecology of these species (Minore 1979), their increase suggests moister conditions and perhaps reduced fire frequency in post-Mazama time. At Davis Lake, south of the glacial maximum, Barnosky (1981) infers increasing moisture beginning at about 5500 yr B.P. from increases in hemlock and cedar. Hebda and Mathewes (1982) have compiled evidence that red cedar appears to reach maximum Holocene abundance at many coastal sites during the last 3000 to 2000 years, a period of increased wetness and cooling that correlates well with other studies. Heusser (1977), in summarizing the palynology of the Pacific slope of Washington, proposes an approximately 1° - 2° decline in mean July temperature during the last 3000 years - a time characterized by a *Tsuga heterophylla-Abies-Thuja* pollen assemblage.

In his earlier review of the late-Pleistocene environments along the Pacific Coast, Heusser (1960, p. 179) also emphasized the role of bog growth as a climatic indicator. The shift from Hypsithermal to cooler late-postglacial conditions was indicated by increases in *Sphagnum* spores and heath (Ericales) plants, suggesting active bog growth beginning around 3000 years ago on the south coast. In a study of Burns Bog on the Fraser River Delta, Hebda (1977) has dated a shift from shrubland to *Sphagnum* peatland at 3000 yr B.P., although the role of climate was not emphasized.

Paleoclimatic reconstructions using pollen-climate transfer functions do not show a late Holocene cooling or increased wetness. Results from Heusser *et al.* (1980) on the Olympic Peninsula indicate cooling after 8000 yr B.P., with a slight rise in temperature in very recent time. Precipitation increased from early Holocene levels, but appears to decrease over the last 2000 years. Mathewes and Heusser's (1981) data (Figure 2) suggest more or less constant conditions over the last 6000 years. Heusser (1983) states that reconstructed precipitation and temperature from a marine core in Saanich Inlet show little change in the Holocene. Thus the paleobotanical evidence is again not as consistent as one would hope.

Interior

In his recent review of palynological data relating to grassland history in the southern interior of British Columbia, Hebda (1982) concluded that modern climatic conditions generally appear around 4500 years ago in response to a cooling trend with increasing moisture. Alley (1976) however, emphasized a change to much cooler and moister conditions in the Okanagan Valley beginning around 6000 yr B.P., with subsequent fluctuations that appear to coincide with neoglacial stades. Most of the sites reviewed by Hebda, in fact, are interpreted as showing a trend to moister conditions (although perhaps still warmer than present) somewhere between 8000 and 6000 yr B.P., with a further moisture increase at about 4500 years ago (Hebda 1982, p. 175). In southeastern British Columbia, for example, Hazell (1979) showed a change from shrubby vegetation in the early Holocene to closed-canopy forest beginning 7000 years ago. Although he did not ascribe climatic significance to the appearance of western hemlock pollen at 4000 yr B.P., this increase of a moisture demanding species probably signifies a climatic shift, in keeping with other data.

A complex post-Mazama climatic history near Lillooet is inferred from detailed stratigraphic, palynological and macrofossil studies in Chilhil and Phair Lakes (King 1980). Increased moisture is indicated by marsh formation at Phair Lake and decreases of mudflat or prairie species at about 7000 years ago. At 5650 yr B.P., a permanent waterbody formed in the Phair Lake basin and marl was laid down. A change to gyttja at 2000 yr B.P., may signal another rise in water level. Modern pollen assemblages, with reduced grass and sage, appear around 4000 years ago, coinciding with an increase in pelecypods and decreasing aquatic gastropods - changes perhaps also related to increasing water levels. At Chilhil Lake, the reductions in grass and sage pollen are radiocarbon-dated at 4400 yr B.P., but water-level changes based on aquatic plant macrofossils and molluscs are not synchronous with Phair Lake. Further investigations of this type are needed to help separate local fluctuations from climatically more useful regional ones.

Some of the potentially most important palynological studies in the southern interior are Hebda's (1982) unpublished sites such as Finney Lake, "Pemberton Hill", and Bluebird

Lake (Figure 1). Full evaluation of their climatic significance must await publication of the pollen diagrams and their geochronologies.

Comparisons of southern British Columbia sites with eastern Washington and western Alberta are also illuminating. According to Hebda (1982), patterns in Washington are similar to those in British Columbia, with evidence for maximum expansion of open terrain between about 9000 and 6000 years ago. Between 5000 and 4000 yr B.P., forest expansion signals a shift to cooler and moister conditions in most areas.

At Chalmer's Bog, southwestern Alberta, vegetation has remained essentially unchanged between 6000 yr B.P. and the present (Mott and Jackson 1982), suggesting the absence of any major late Holocene climatic shifts. In the same area, Harris and Pip (1973) concluded from fossil molluscs that cooling may have occurred at around 7000 yr B.P., with little change thereafter. White and Mathewes (1982) and Vance *et al.* (1982) review other evidence from Alberta and northeastern B.C. which indicates a period of maximum warmth and aridity prior to about 6,000 years ago, followed by different episodes of cooling. Vance *et al.* (1983) place the end of the Hypsithermal at 4000 yr B.P., and the advent of modern forest conditions at 3000 years ago.

The general patterns of early to mid-Holocene warm/dry conditions, followed by a cooling trend culminating around 4000 yr B.P., suggest the operation of macroclimatic factors modified by local site differences. The current trend toward using lake cores for analysis coupled with radiocarbon dating and tephrochronology is a good one. Perhaps a clearer concensus will emerge in the near future about the true nature of Holocene climates in western Canada.

SUMMARY

(1) Few published paleobotanical data relating to climate are available for the period prior to 12,000 years ago. Tentative suggestions of more continental conditions (colder and perhaps somewhat drier than present) have been made for coastal and interior British Columbia.

(2) Generally cool and moist conditions are proposed from about 12,000 to 10,500 years ago. A rapid climatic amelioration between about 10,500 and 10,000 yr B.P. is inferred at several coastal sites.

(3) An early Holocene period of climate warmer and drier than present is inferred from many sites prior to the Mazama ashfall of about 6600 years ago. Peak xerothermic conditions appear to end around 7500 to 7000 yr B.P. at most south-coastal and interior sites. This interpretation is based on both qualitative and numerical reconstructions using transfer functions.

(4) A poorly-defined period of climatic transition characterizes the mid-Holocene. A gradual cooling and wettening trend is suggested after 7000 yr B.P., resulting in higher pond and lake levels, tree encroachment on previously more open terrain, and range expansion of trees requiring higher moisture. Interior sites are particularly variable, and the nature and timing of these events are not uniform. The end of the so-called Hypsithermal is time-transgressive.

(5) By 4500 - 3000 yr B.P., relatively modern conditions appear in most regions. Another climatic deterioration is suggested at many sites, correlative with neoglacial activity, although numerical transfer functions on the coast do not support this interpretation.

(6) These conclusions must remain tentative until final publication of some of the reviewed data, and until more supporting data from disciplines other than paleobotany are available. Data from animal fossils and isotopic temperature studies are necessary to test the paleobotanical inferences, which are only proxy data for climatic change.

REFERENCES

Alley, N.F. 1976. The palynology and paleoclimatic significance of a dated core of Holocene peat, Okanagan Valley, southern British Columbia. Canadian Journal of Earth Sciences 13:1131-1144.

Barnosky, C.W. 1981. A record of late Quaternary vegetation from Davis Lake, southern Puget Lowland, Washington. Quaternary Research 16:221-239.

Birks, H.J.B. 1981. The use of pollen analysis in the reconstruction of past climates: a review. In: Climate and History. Edited by: T.M.L. Wigley, M.J. Ingram and G. Farmer. Cambridge University Press. pp. 111-138.

Clague, J.J. 1981. Late Quaternary geology and geochronology of British Columbia: Part 2: Summary and discussion of radiocarbon-dated Quaternary history. Geological Survey of Canada Paper 80-35:1-41.

Conway, E. 1957. Spore production in bracken (*Pteridium aquilinum* (L.) Kuhn). Journal of Ecology 45:273-284.

Courtin, P.J., K. Klinka, J. Heineman, W. Mitchell, and R.N. Green. 1981. Biogeoclimatic units, Victoria-Vancouver (MAP) 1:500,000 NTS Sheet 92 S.E. British Columbia Ministry of Forests, Vancouver.

Davis, M.B. 1973. Pollen evidence of changing land use around the shores of Lake Washington. Northwest Science 47:133-148.

______. 1976. Pleistocene biogeography of temperate deciduous forests. Geoscience and Man 13:13-26.

______. 1978. Climatic interpretation of pollen in Quaternary sediments. In: Biology and Quaternary Environments. Edited by: D. Walker and J.C. Guppy. Australian Academy of Sciences. pp. 35-51.

Deevey, E.S., and R.F. Flint. 1957. Postglacial Hypsithermal interval. Science 125:182-184.

Hansen, B.S., and D.J. Easterbrook. 1974. Stratigraphy and palynology of late Quaternary sediments in the Puget Lowland, Washington. Geological Society of America Bulletin 85: 587-602.

Hansen, H.P. 1940. Paleocology of two peat bogs in southwestern British Columbia. American Journal of Botany 27:144-149.

______. 1947. Climate versus fire and soil as factors in postglacial forest succession in the Puget Lowland of Washington. American Journal of Science 245:265-286.

______. 1950. Pollen analysis of three bogs on Vancouver Island, Canada. Ecology 38: 270-276.

______. 1955. Postglacial forests in south central and central British Columbia. American Journal of Science 253(11):640-658.

______. 1967. Chronology of postglacial pollen profiles in the Pacific Northwest (U.S.A.). Review of Palaeobotany and Palynology 4:103-105.

______. 1974. Pollen analysis of a deep section from Lake Washington, Seattle. Birbal Sahni Institute of Paleobotany, Special Publication 5:1-4.

Harris, S.A., and E. Pip. 1973. Molluscs as indicators of late and post-glacial climatic history in Alberta. Canadian Journal of Zoology 51:209-215.

Hazell, S.D. 1979. Late Quaternary vegetation and climate of Dunbar Valley, British Columbia. M.Sc. thesis, University of Toronto.

Hebda, R. 1977. The paleoecology of a raised bog and associated deltaic sediments of the Fraser River Delta. Ph.D. thesis, University of British Columbia, Vancouver.

______. 1982. Postglacial history of grasslands of southern British Columbia and adjacent regions. In: Grassland Ecology and Classification; Symposium Proceedings. British Columbia Ministry of Forests. Edited by: A.C. Nicholson, A. McLean, and T.E. Baker. pp.157-191.

______. (in press) Vegetation history at Bear Cove Bog, near Port Hardy, Vancouver Island, B.C. Canadian Journal of Botany.

Hebda, R.J., and R.W. Mathewes. 1982. Pollen record of cedar and development of native cultures of the North Pacific Coast. 35th Annual Northwest Anthropological Conference, Simon Fraser University, Burnaby. Abstract.

Heusser, C.J. 1960. Late-Pleistocene environments of North Pacific North America. American Geographical Society Special Publication No. 35:1-308.

______. 1973. Environmental sequence following the Fraser advance of the Juan de Fuca Lake, Washington. Quaternary Research 3:284-306.

______. 1974. Quaternary vegetation, climate, and glaciation of the Hoh River Valley, Washington. Geological Society of America Bulletin 85:1547-1560.

______. 1977. Quaternary palynology of the Pacific slope of Washington. Quaternary Research 8:282-306.

______. 1978. Modern pollen rain of Washington. Canadian Journal of Botany 56:1510-1517.

Heusser, C.J., L.E. Heusser, and S.S. Streeter. 1980. Quaternary temperatures and precipitation for the north-west coast of North America. Nature 286:702-704.

Heusser, L.E. 1983. Palynology and paleocology of postglacial sediments in an anoxic basin, Saanich Inlet, British Columbia. Canadian Journal of Earth Sciences 20:873-885.

Hills, L.V., and E.V. Sangster. 1980. A review of paleobotanical studies dealing with the last 20,000 years; Alaska, Canada, and Greenland. In: Climatic Change in Canada. Edited by: C.R. Harington. Syllogeus No. 26:73-224.

King, M. 1980. Palynological and macrofossil analyses of lake sediment from the Lillooet area, British Columbia. M.Sc. thesis, Simon Fraser University, Burnaby.

Leopold, E.B., R. Nickman, J.I. Hedges, and J.R. Ertel. 1982. Pollen and lignin records of late Quaternary vegetation, Lake Washington. Science 218:1305-1307.

Mathewes, R.W. 1973. A palynological study of postglacial vegetation changes in the University Research Forest, southwestern British Columbia. Canadian Journal of Botany 51:2085-2103.

Mathewes, R.W., and J. D'Auria. 1982. Historic changes in an urban watershed determined by pollen and geochemical analyses of lake sediment. Canadian Journal of Earth Sciences 19:2114-2125.

Mathewes, R.W., and L.E. Heusser. 1981. A 12,000-year palynological record of temperature and precipitation trends in southwestern British Columbia. Canadian Journal of Botany 59:707-710.

Mathewes, R.W., and G.E. Rouse. 1975. Palynology and paleoecology of postglacial sediments from the lower Fraser River Canyon of British Columbia. Canadian Journal of Earth Sciences 12:745-756.

McLennan, D.S. 1981. Pollen transport and representation in the Coast Mountains of British Columbia. M.Sc. thesis, Simon Fraser University, Burnaby.

McMinn, R.G. 1951. The vegetation of a burn near Blaney Lake, British Columbia. Ecology 32:135-140.

Minore, D. 1979. Comparative autecological characteristics of northwestern tree species: a literature review. Pacific Northwest Forest and Range Experiment Station, United States Department of Agriculture, Forest Service, General Technical Report PNW-87.

Mott, R.J., and L.E. Jackson, Jr. 1982. An 18,000 year palynological record from the southern Alberta segment of the classical Wisconsinan "Ice-free Corridor". Canadian Journal of Earth Sciences 19:504-513.

Page, C.N. 1976. The taxonomy and phytogeography of bracken - a review. Botanical Journal of the Linnaean Society 73:1-34.

Terasmae, J., and J.G. Fyles. 1959. Palaeobotanical study of late-glacial deposits from Vancouver Island, British Columbia. Canadian Journal of Botany 37(5):815-817.

Vance, R.E., D. Emerson, and T. Habgood. 1983. A mid-Holocene record of vegetative change in Central Alberta. Canadian Journal of Earth Sciences 20:364-376.

Watt, A.S. 1976. The ecological status of bracken. Botanical Journal of the Linnaean Society 73:217-239.

Webb, T., III. 1980. The reconstruction of climatic sequences from botanical data. Journal of Interdisciplinary History 10:749-772.

White, J.M. 1983. Late Quaternary geochronology and palaeoecology of the Upper Peace River District, Canada. Ph.D. thesis, Simon Fraser University, Burnaby.

White, J.M., and R.W. Mathewes. 1982. Holocene vegetation and climatic change in the Peace River District, Canada. Canadian Journal of Earth Sciences 19:555-570.

Wright, H.E., Jr. 1966. Stratigraphy of lake sediments and the precision of the palaeocolimatic record. In: World Climate from 8,000 to 0 B.C. Edited by: J.S. Sawyer. Royal Meteorological Society, London. pp. 157-173.

______. 1976. The dynamic nature of Holocene vegetation: a problem in paleoclimatology, biogeography, and stratigraphic nomenclature. Quaternary Research 6:581-596.

RECONSTRUCTION OF ENVIRONMENTAL CONDITIONS IN THE EASTERN CANADIAN ARCTIC DURING THE LAST 11,000 YEARS

John T. Andrews[1]

INTRODUCTION

The validity and accuracy of Quaternary environmental reconstructions is limited by: (1) the nature of the stratigraphic record, i.e. either continuous or discontinuous sedimentation; (2) the fidelity of the system being queried to record a specific shift in one or more environmental fluctuations; and (3) the accuracy and precision of the dating method. Within the eastern Canadian Arctic discontinuous and continuous sedimentary sequences exist, although by far the bulk of past research has focused on dating discontinuous stratigraphic sequences, such as segments of raised marine strata. Only within the last few years has a consistent effort been made to obtain lake or marine cores with the aim of providing continuous stratigraphic records.

The fidelity of a system to record an environmental change depends partly on the typical relaxation time that a system has to a specific perturbation. In part the response is probably a function of both the amplitude and duration of the event. How well we can interpret such events is largely a matter of the sedimentation rate within an environment. Table 1 lists some of the situations in the eastern Canadian Arctic where continuous records have been obtained and outlines some limits on sedimentation rates expressed as the number of years to accumulate a centimetre of material (yr cm^{-1}).

Dating periods of climatic change in arctic areas has relied very heavily on radiocarbon dating and on lichenometry. Radiocarbon dating organic or carbonate materials of < 11,000 yr B.P. is not divorced from a variety of problems (e.g. Blake 1975; Stuckenrath *et al.* 1979; Fillon and Harmes 1982). Lichenometric age curves (e.g. Miller 1973; Andrews and Barnett 1979) are in part controlled by radiocarbon dates and have other errors associated with age determinations (Locke *et al.* 1979). The combination of sedimentation rate and dating accuracy and precision together act as a filter which removes a specific frequency

---

[1] INSTAAR and Department of Geological Sciences, University of Colorado, Boulder, Colorado 80309

TABLE 1: *SEDIMENTATION AND ACCUMULATION RATES IN THE EASTERN CANADIAN ARCTIC (SEE FIGURE 5).*

MATERIAL	TIME SPAN[1] (ka)	SEDIMENTATION RATE (yrs/cm)	ACCUMULATION RATE (g/cm^2/yr)
Snow	10 - 70	.03	30
Fiord muds	10 - (>100)?	.01 - 60	100 - .017
Lake sediments	6 - (70)?	50 - 100	.02 - .01
Peats	2 - 9	20 - 100	.02 - .004
Deep-sea muds	2 - >200	100 - 240	.012 - .005

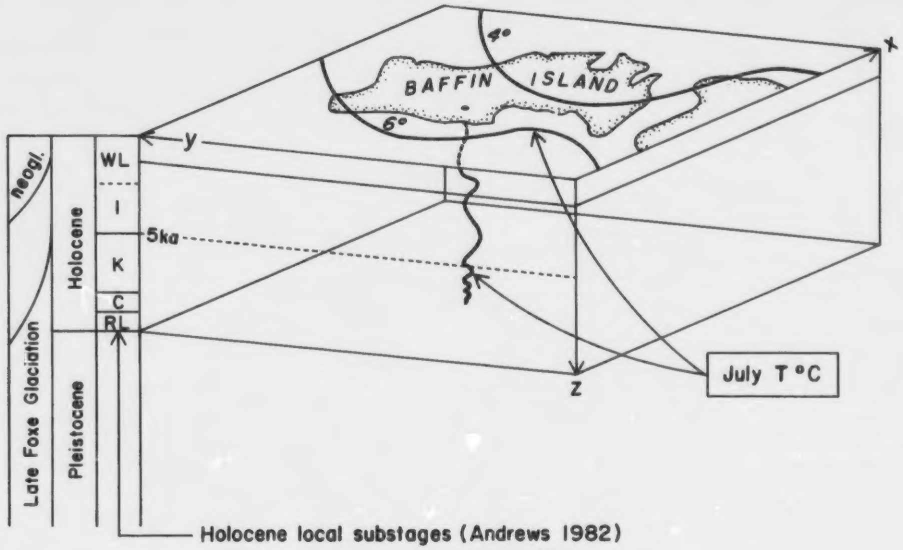
[1] Ages bracketed and followed by a question mark have not been proven, but are judged possible.

from any single record. Additional filtering takes place through the sampling demands of a technique. For example, paleomagnetic analysis tends to be limited to a 3-cm sampling interval because of the disruption of sediment around the 1-cc sampling cube. In other instances the sampling interval appears to be largely a historical response - thus many laboratories sample at X cm regardless of changes in the rate of sedimentation!

This paper has two broad objectives: the *first* is to discuss briefly attributes of available Holocene climatic time-series from Baffin Island; the *second* is to present and discuss a series of maps in which I attempt to portray changing conditions over the last 11,000 years. These two approaches represent different facets of climatic reconstructions. They are not mutually exclusive, but rather need to be used together to verify or falsify particular hypotheses that have been derived from a single site or for a region. Thus we are concerned with a three-dimensional view in which the x and y axes represent geographic space whereas the z axis represents time (Figure 1). Quaternary scientists have traditionally been concerned with a two-dimensional view of their universe, x/y co-ordinates being reduced to a single point. Largely as the result of the CLIMAP Project (Cline and Hays 1976) and its various son and daughter projects, attention has now been directed toward the reconstruction of environments at particular moments of time. The larger part of this paper attempts to ascertain how successfully this can be accomplished for the eastern Canadian Arctic, or more specifically, the area from northernmost Labrador to northern Baffin Island and from Baffin Bay to Foxe Basin (Figure 2).

HOLOCENE TEMPERATURE RECORDS

Figure 3 shows three temperature series. The first is for the modern July T°C at the settlement of Clyde River, on the outer coast of Baffin Island facing Baffin Bay. The second record is an estimate of July T°C derived from an analysis of the pollen content of a peat site at Windy Lake, Pangnirtung Pass (Andrews *et al.* 1980; Davis 1980), and the third time-series is also derived from pollen analysis, but in this case it is from lake sediments at Iglutalik Lake, outer Pangnirtung Fiord (Davis 1980; Andrews *et al.* 1980). The Clyde River data are based on mean daily measured temperatures. The other records, as noted, were derived from transfer-function equations (in this case using stepwise linear regression) in which the present pollen rain is associated with the present climate in the form: climate=f (pollen) (Webb and Bryson 1972; Andrews *et al.* 1980). The Windy Lake series


x
BAFFIN ISLAND
4°
6°
y
neogl.
Late Foxe Glaciation
Holocene
Pleistocene
WL
I
K
C
RL
5ka
z
July T °C

Holocene local substages (Andrews 1982)

		Duration ^{14}yrs.
WL	= Windy Lake	0 – 2.5
I	= Iglutalik Lake	2.5–5
K	= Kangilo	5 – 8
C	= Cockburn	8 – 9
RL	= Remote Lake	9 – 10ka

FIGURE 1: *Diagram showing Baffin Island July temperature reconstructions in space and time. The subdivision for the Holocene is taken from Andrews and Ives (1978) and Andrews (1982).*

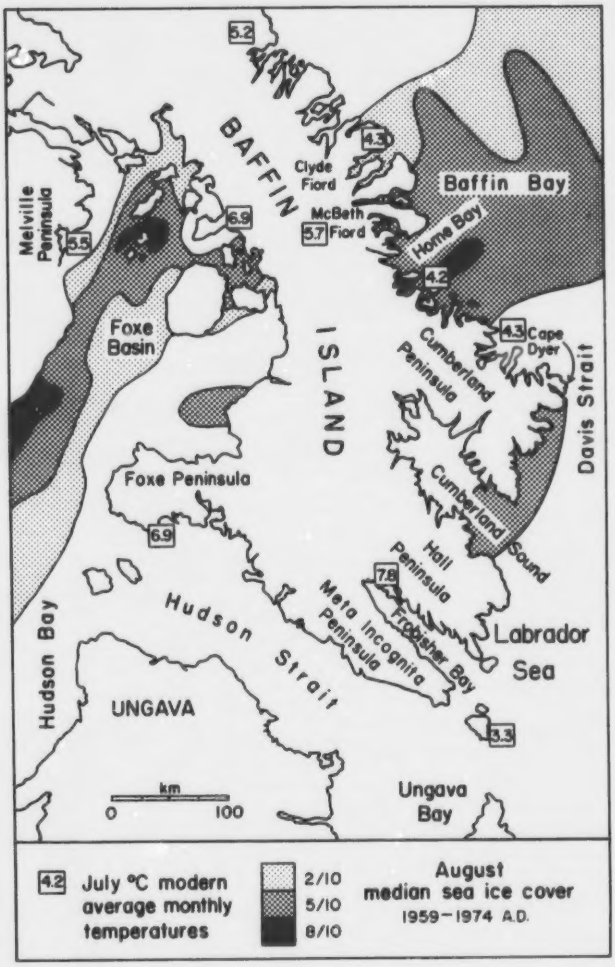

5.2
BAFFIN
ISLAND
4.3
Clyde Fiord
Baffin Bay
McBeth Fiord
5.7
6.9
Melville Peninsula
5.5
Home Bay
4.2
Foxe Basin
4.3
Cape Dyer
Cumberland Peninsula
Davis Strait
Cumberland Sound
Foxe Peninsula
6.9
Hall Peninsula
7.8
Meta Incognita Peninsula
Frobisher Bay
Labrador Sea
Hudson Bay
Hudson Strait
UNGAVA
3.3
km
0
100
Ungava Bay
4.2 July °C modern average monthly temperatures
2/10
5/10
8/10
August median sea ice cover 1959–1974 A.D.

FIGURE 2: *Map of present summer conditions in the eastern Canadian Arctic. July temperatures are taken from the* Canadian Weather Review *and the median ice conditions from the atlas of Markham (1981).*

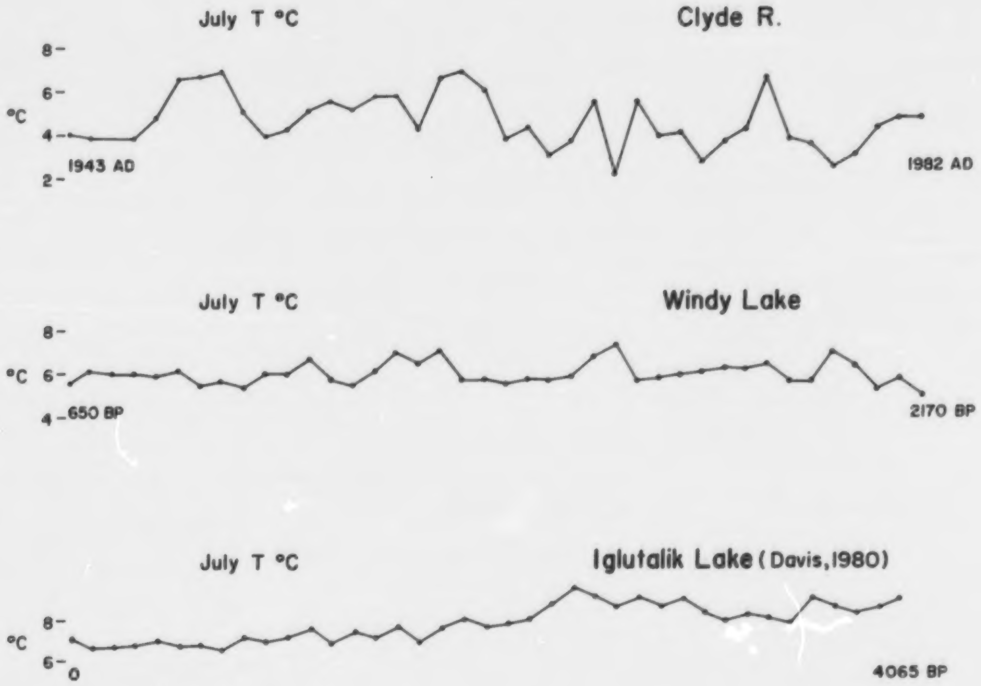


FIGURE 3: *July temperatures for Clyde River (measured) and for Windy Lake and Iglutalik Lake (predicted) (Andrews* et al. *1980; Davis 1980).*

has one temperature estimate for approximately every 40 years, whereas the Iglutalik Lake series has one estimate every 105 years. The derivation of the transfer functions is discussed in detail in Andrews *et al.* (1980) and applied in Davis (1980), Andrews *et al.* (1981A), and Short and Jacobs (1982).

Figure 4 illustrates the power spectra of these three series. This is presented largely for illustrative purposes as the treatment to date has not been very rigorous. On the basis of exotic pollen peaks in the Windy Lake peat, Nichols *et al.* (1978) suggested that there was a 250-year periodicity to the pollen rainout of taxa such as *Picea, Pinus,* and *Alnus*.

Figures 3 and 4 indicate that the variance of the modern time-series is larger than estimated from the late Holocene sites. The high resolution and low resolution spectra for Windy Lake for n=40 and n=89 (Figure 4) show some differences, but whether these are associated with real temperature changes over the different spans of time remains to be tested (e.g. Figure 3). I have no answer to the problem, but the question has to be posed. How do we compare records that have temporal resolutions in the ratios of 1:40:105? Does our July temperature estimate from Windy Lake represent an integrated 40-year average, or does it represent a single yearly estimate within a 40-year span? Comparison of the Windy Lake and Iglutalik Lake temperature estimates over the late Holocene (see Andrews *et al.* 1981A) suggests a close correspondence between the lake and peat records. In the time domain a significant deterioration of climate occurred about 2000 years ago (Figure 3).

On Baffin Island, and elsewhere in arctic Canada, there are currently few detailed time-series that span much of the Holocene. I suggest that for climatic interpretations the sedimentation rate of the system should approach 5-10 yr/cm. This resolution is exceeded in studies of ice cap stratigraphy (Koerner and Fisher 1981), and is probably exceeded in fiords where sedimentation rates close to ice fronts may be close to 0.03 yr/cm (Gilbert 1982). However, in such situations high rates of inorganic clastic inputs may mask the biological signal that is traditionally used to interpret environmental changes.

I have not discussed, in terms of Figures 3 and 4, the problems associated with calibrating the response of a system to environmental change. Thus, there is a real difference between calibrating pollen/climate in terms of "modern" conditions and evaluating the sensitivity of the vegetation/pollen response to climatic change.

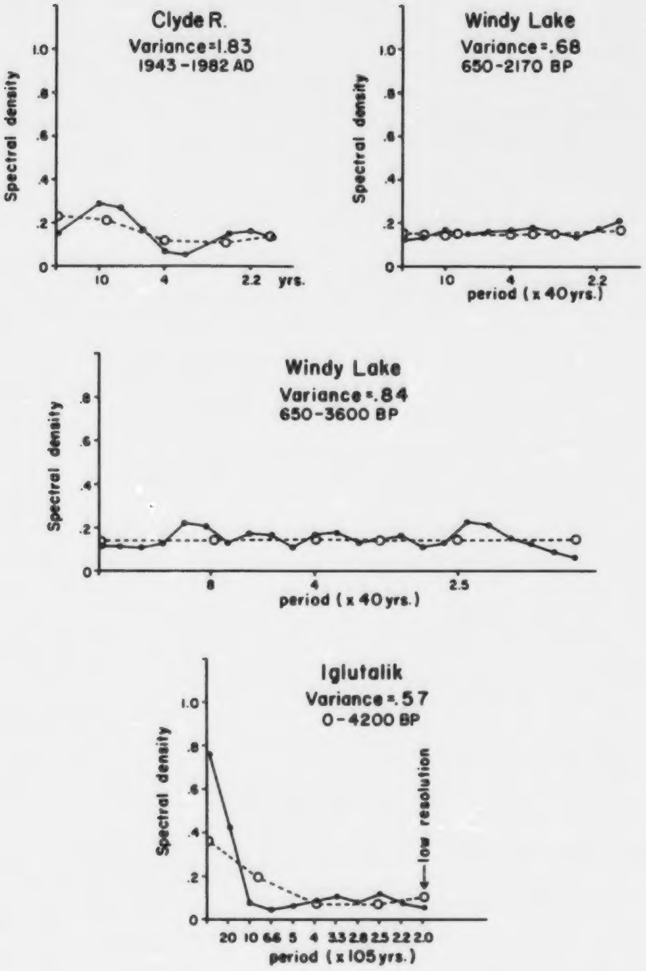


FIGURE 4: *Power spectra for the time series of Figure 3.*

ENVIRONMENTAL CHANGES DURING THE LAST 11,000 YEARS

In the eastern Canadian Arctic we have a variety of sediment traps that should provide continuous records covering the Holocene and latest Pleistocene (Figure 5). However, these different situations have not been explored in great depth. Lake sediments have only been obtained in the last few years (e.g. Mode 1980; Davis 1980); ice core records for the Holocene have yet to be obtained for detailed study, and nearshore (fiord and shelf) sediments are only now being examined (e.g. Osterman 1982).

On Figures 6-10, I have mapped our best estimates of several parameters: (1) the extent of glacial ice; (2) the severity of the inshore marine environment; and (3) July temperature estimates based on radiocarbon-dated and pollen-calibrated terrestrial sites. Figure 2 represents the present-day situation, whereas Figures 6-10 represent "snapshots" of the eastern Canadian Arctic at about 3000, 6800, 8000, 10,000 and 10,700 radiocarbon years ago. This is the first time I have tried to incorporate a variety of field and laboratory data into a series of pictures of the eastern arctic environment. So Figures 6-10 are best considered a series of reasonable hypotheses that may provide a framework for future research. In the remainder of this paper I will discuss and comment on each map.

PRESENT DAY (FIGURE 2)

This map reports present average July temperatures for weather stations on Baffin Island, and average sea-ice conditions for late August. There is a decided lag between the terrestrial "high summer" and the marine equivalent of the order of 1-2 months. Sources for the information on Figure 2 are listed in the figure caption. But how different were past conditions from the present?

Average July temperatures over Baffin Island are based on a variety of records, several of which cover different intervals of time. The longest records, such as Clyde River and Frobisher Bay, date from the early to mid-1940s, whereas several other sites are associated with DEW Line stations and commenced keeping weather records in the middle to late 1950s.

Figure 2 shows that there is a decided west to east decrease in average July temperature. The coldest reporting weather station was on Resolution Island at the mouth of Hudson Strait, where average July temperature is only 3.3°C. At the settlement of Frobisher Bay, however, some 200 km to the northwest, average July temperature is 7.8°C. Fort Chimo

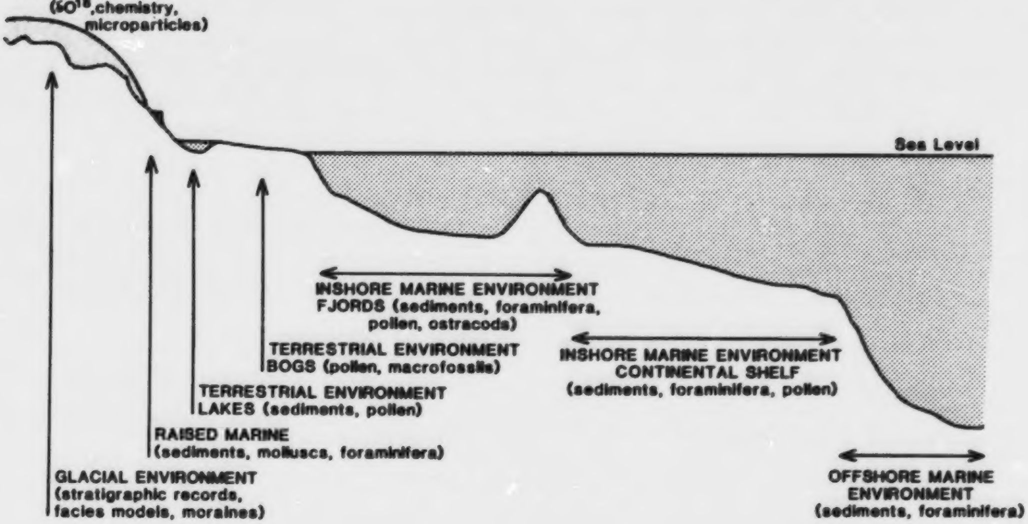

ICE CAP
(δO^{18},chemistry,
microparticles)
Sea Level
INSHORE MARINE ENVIRONMENT
FJORDS (sediments, foraminifera,
pollen, ostracods)
TERRESTRIAL ENVIRONMENT
BOGS (pollen, macrofossils)
INSHORE MARINE ENVIRONMENT
CONTINENTAL SHELF
(sediments, foraminifera, pollen)
TERRESTRIAL ENVIRONMENT
LAKES (sediments, pollen)
RAISED MARINE
(sediments, molluscs, foraminifera)
GLACIAL ENVIRONMENT
(stratigraphic records,
facies models, moraines)
OFFSHORE MARINE
ENVIRONMENT
(sediments, foraminifera)

FIGURE 5: *Depositional environments in the eastern Canadian Arctic that may provide continuous stratigraphic sequences.*

at the southern end of Ungava Bay, by comparison, has an average July temperature of about 10°C.

Along the eastern coast of central and northern Baffin Island, July temperatures are remarkably consistent and are close to 4.3°C (Figure 2). The land surrounding Foxe Basin has a pattern of July temperature suggesting an east to west decrease across the Basin. Thus, average temperatures at Hall Beach are 5.5°C, whereas on the eastern side of Foxe Basin the Longstaff Bluff DEW Line Station has an average of 6.9°C.

The pattern of "high summer" temperatures is strongly associated with the cooling effect of extensive sea-ice cover in both Foxe Basin and in Baffin Bay/northern Labrador Sea. On Baffin Island, fiord heads do not generally become ice free until mid-July, and the outer east coast can have heavy ice offshore through late August and on into September (Figure 2).

Figure 2 also illustrates the median cover of sea ice during late August for the period 1959-1974 (Markham 1981). The extremes for the median condition can be very large (see Crane 1978; Jacobs and Newell 1979). Ice clears from Baffin Bay both from the east and north, in response to warm waters moving northward and then southward, as part of the West Greenland Current and as the North Water expands. The pattern of clearing and the set of the coast explains why Home Bay is the last area clear of heavy sea ice. Within Foxe Basin, heavy sea ice (up to 8/10th's cover) persists in the central part and along the west coast until September. In many years, Home Bay and Foxe Basin never become totally ice free, and small boat travel inshore can be very slow and hazardous.

On land, low elevation sites are largely snow-free by mid-June, and by mid-to-late July, in an average summer, the transitional snowline is close to the elevation of glacier equilibrium lines (ELA's). Andrews and Miller (1972, Figures 1A, B) mapped the lowest ELA across Baffin Island using glacier morphology as a guide to the long-term equilibrium altitude (e.g. Andrews 1975). Small cirque and valley glaciers have ELA's of 300-400m off southeastern Baffin Island and between 500-600m in outer Home Bay. There is a gradient inland of glacier ELA's, so that along the fiord heads mapped altitudes rise to between 700 and 900 m above sea level (asl).

The topography of Baffin Island is such (Ives 1962; Ives *et al.* 1975; Andrews *et al.* 1976; Locke and Locke 1977; Williams 1978) that a slight decrease in the regional snowline elevation can cause a restricted advance of mountain glaciers, whereas on the high uplands of central Baffin Island a similar fall of snowline causes a massive extension of permanent snowcover. Radiocarbon dates on dead moss and lichens suggest that during the

Little Ice Age there was a considerable expansion of permanent snow above 500-600 m asl. The warming of the last several decades has resulted in many of these snowfields disappearing (Falconer 1962), leaving light-toned, vegetation-free rock surfaces that are easily visible on 1:1,000,000 LANDSAT imagery (Andrews *et al.* 1976). Many, but not all glaciers show some evidence of recession over the last few decades (Falconer 1962; Andrews *et al.* 1970; Miller 1973; Locke 1980; Davis 1980).

One brief point before leaving present and looking at past conditions - in many mountain areas of eastern Baffin Island it is quite apparent that the Little Ice Age (or neoglaciation) represents an expansion of these ice bodies to dimensions that *exceed* those reached during the late Wisconsin (called the late Foxe Glaciation on Baffin Island) Glaciation. This was first reported by Rothlisberger (1951), and has been mentioned more recently by Harrison (1964), Miller (1976), and Locke (1980). These conditions do not prevail everywhere, but it is especially clear in the mountains near Clyde River and Cape Dyer. This situation is analagous to the conditions in Antarctica today where, in places, alpine glaciers are extending across deposits left by the late Pleistocene retreat of the ice sheet (Stuiver *et al.* 1981).

APPROXIMATELY 10,700 YEARS AGO (LATEST PLEISTOCENE, FIGURE 6)

Environmental conditions on Baffin Island and in Baffin Bay prior to the Pleistocene/Holocene boundary at 10,000 years ago are difficult to reconstruct. Very few sites on eastern Baffin Island are known to have shells or peats of this age. However, work on deep sea cores (Aksu 1981; Fillon and Duplessy 1980), and work in progress on cores on the Baffin Island shelf, will eventually shed more light on conditions that prevailed. Cores now being studied are shown on Figure 6 (the legend for that map applies to the other maps too).

The extent of late Foxe (=Wisconsin) glacial cover over eastern Baffin Island is controversial (e.g. Denton and Hughes 1981; Miller and Dyke 1974; Andrews 1980; Dyke *et al.* 1982), but field workers on Baffin Island have used a variety of relative and "absolute" dating methods to infer that the Laurentide ice sheet complex did not extend onto the continental shelf, except off Hudson Strait and Frobisher Bay (Osterman 1982; Osterman *et al.* in press; Stravers and Miller 1983). New radiocarbon dates on organic carbon in the marine cores located on Figure 6 apparently support the evidence for a relatively

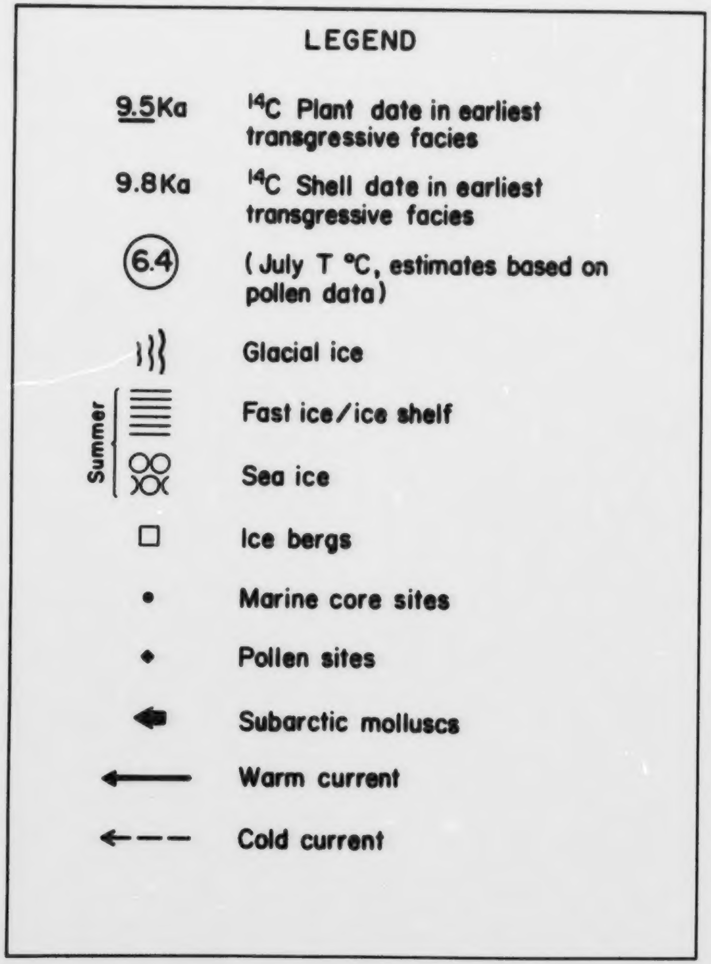

LEGEND
9.5Ka 14C Plant date in earliest transgressive facies
9.8Ka 14C Shell date in earliest transgressive facies
6.4 (July T °C, estimates based on pollen data)
Glacial ice
Summer
Fast ice/ice shelf
Sea ice
Ice bergs
Marine core sites
Pollen sites
Subarctic molluscs
Warm current
Cold current

FIGURE 6A: *Legend for use in Figures 6 - 10.*

10.7 Ka.

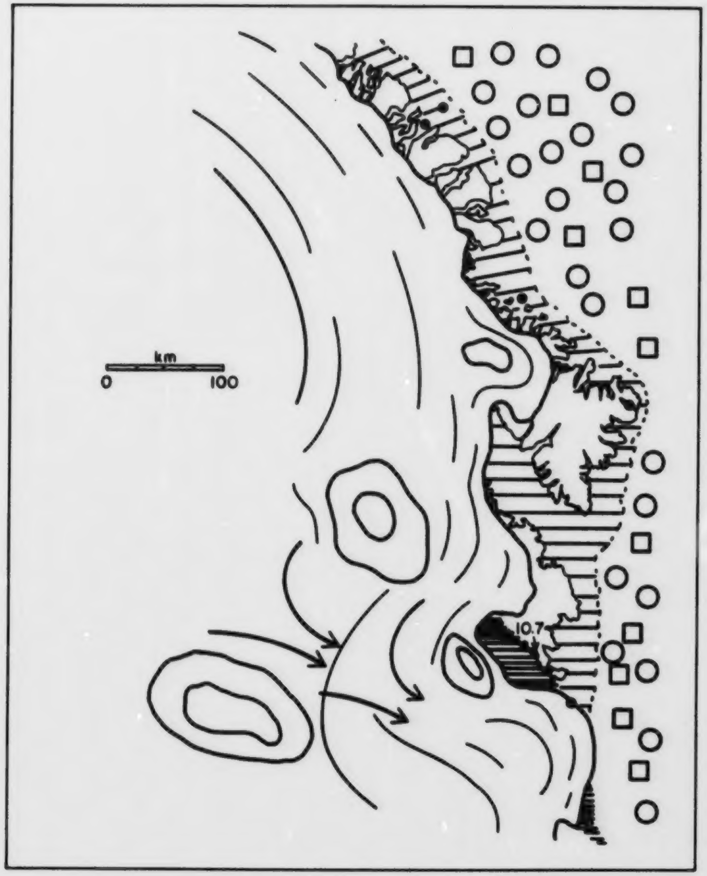


FIGURE 6B: *The eastern Canadian Arctic about 10,700 years ago.*

limited late Foxe glacial maximum, however, it must be stressed that there are reasons to question radiocarbon dates on organic carbon. Some dates may well be accurate, but all such dates are probably best prefixed by a < or = symbol. One core off Cape Aston, Baffin Island, in 99m of water has a date on marine mollusc shells of approximately 11,000 yr B.P. (Nelson, in Andrews and Short, in press).

On land the only date for this interval is a collection of shells reported in Miller (1980). The shells come from a deposit which lies at the terminus of the Hall readvance. Beyond the limits of this event Miller (1980) noted an older drift which was considerably more weathered; thus he concluded that the Hall moraines mark the outer limit of glacial ice in Frobisher Bay during the late Foxe Glaciation.

Figure 6 poses a significant question: If there were extensive ice-free areas during late Foxe time, why have no deposits been located that date from this interval (other than the Hall deposits)? Remember that on Baffin Island, like most areas in the Canadian Arctic, the date of deglaciation is often deduced from the stratigraphy and age of associated glacial marine and marine strata (Craig and Fyles 1960; Andrews 1970; Blake 1970; England 1976). However, Andrews (1975), Dyke *et al.* (1982), and England (1983) have noted that raised marine deposits do occur beyond the physical limits of glaciation, and that there is evidence for sea-level stability and even transgressions during the latest Pleistocene/earliest Holocene.

Two answers have been suggested to the above question. Andrews (1980) suggested that, prior to about 10,000 years ago, relative sea level off Baffin Island was below present due to a combination of glacial isostatic depression and a lower global sea level. However, Quinlan (1981) has attempted to model the relative sea level history of the eastern Canadian Arctic using a realistic earth rheological model (e.g. Quinlan and Beaumont 1982; Peltier and Andrews 1976). Quinlan's results suggest that relative sea level was *above* present prior to 10,000 years ago. This result was obtained with an ice sheet reconstruction not unlike Figure 6, i.e. ice did *not* extend to the outer coast. If Quinlan's reconstruction is correct, it suggests that some other process has to be invoked in order to cause a dearth of sedimentary sequences associated with higher sea levels.

Figure 6 suggests that a fringing ice shelf, much like that which exists today along the coast of northern Ellesmere Island (Crary 1960; Lyons and Mielke 1973) extended along the outer coast of Baffin Island. Its outer limit probably coincided with the shelf break and thus had dimensions very similar to the present fast-ice distribution. I envisage the shelf extending inland along the fiords where it would buttress fiord glaciers flowing seaward from the Laurentide ice sheet complex, or from local centres such as the Penny Ice Cap. Under present conditions, the coastal fast ice breaks up in middle July to late August, although in some years (such as the infamous 1972!; see Figure 3) the fast ice in Home Bay was not dislodged. This suggests that we can take the 1972 summer as a *minimum* climatic analogue for the environment portrayed on Figure 6.

During the summer of 1972, mean monthly temperatures were consistently 1°-3°C below average (Table 2) thus reducing the overall average summer temperature at reporting stations considerably. If these, or conditions of greater severity prevailed for several summers in a row, the fast-ice sheet would thicken to such an extent that it probably would not disappear during the "normal" summer. Table 2 also lists the climate of Alert, northern Ellesmere Island - the nearest weather station to the Ward Hunt ice shelf - as another possible climatic analogue. Mercer (1983, p.118) notes that "..However, ice shelves are now absent from those parts of the Antarctic Peninsula where midsummer temperatures are above the freezing point, probably because such temperatures bring the temperature of the sea surface above -1.5°C."

In Frobisher Bay, Osterman (1982) has studied core HU77-159, which lies in about 500+m of water in the middle of the Bay. Osterman interpreted the sediments and foraminifera in the lower part of the core as indicating a short-lived ice-shelf episode. The core is radiocarbon dated on organic carbon, but the inferred timing of events, such as deglaciation, is in agreement with shell dates in raised marine sediments (Miller 1980; Osterman 1982).

If the scenario of Figure 6 is correct, this would predict that offshore marine cores from the fiords and shelf of Baffin Island should contain a "barren" zone dated about 10,700 years ago. During this period ice rafting of sand-sized particles from iceberg sources would be drastically curtailed as icebergs would be held offshore by the shelf (Figure 6).

10,000 YEARS AGO (REMOTE LAKE SUBSTAGE, FIGURE 7)

A dramatic change must have occurred between 10,700 - 11,000 yr B.P. and 10,100-9,500 yr B.P. (Figure 7). Suddenly, along the outer coast of eastern Baffin Island sediments containing in situ marine faunas are found at elevations up to about 25 m on the outer coast and up to 70+m in Frobisher Bay (Andrews 1975; Miller 1980; Nelson 1982). At other

TABLE 2: *CLIMATE OF EASTERN CANADIAN ARCTIC DURING THE SUMMER OF 1972 (^oC)*[1].

STATION	MAY	JUNE	JULY	AUGUST	SUMMER	SEPTEMBER
Alert	-11.1(0)	-2.2(-2.2)	1.7(-2.2)	0.6(0)	0	-10(-0.6)
Clyde River	-9.4(-2.8)	-2.2(-3.3)	2.8(-2.2)	2.2(-1.7)	0.9	-1.7(-1.7)
Cape Dyer	-6.7(0)	-2.2(-2.8)	3.3(-2.8)	2.8(-2.2)	1.3	-2.8(-1.7)
Frobisher Bay	-6.7(-3.3)	0.0(-3.9)	6.1(-1.7)	5(-1.7)	3.7	1.1(-1.1)
Resolution Island	-6.7(-3.9)	-1.1(-2.2)	2.2(-1.1)	1.7(-1.7)	0.9	-1.1(-2.8)

[1] Figures in brackets refer to the difference between the long term average and the average for 1972.

localities, radiocarbon dates on peaty soils come from the same time interval and underlie marine transgressive facies (Nelson 1982; Mode 1980). Obviously, seasonal open water conditions occurred at this moment in time. Marine molluscs collected from these exposures have a Pan-Arctic fauna that includes no Subarctic molluscs. The fauna is dominated by, and largely limited to: *Mya truncata* and *Hiatella arctica*. The collecting sites are such that the shells were living in relatively shallow water - between 40 and 10 m deep.

I suggest (Figure 7) that break up of the ice shelf/fast ice sheet (Figure 6) resulted in enhanced iceberg drifts along the coast. Because the Laurentide ice sheet was still massive (Figure 7), the crust was still depressed and relative sea level was 70-25 m *above* present, thus allowing slightly deeper draughted icebergs to track more toward the coast and possibly into the fiords.

What might the effect of the ice-shelf disintegration have been on fiord glaciers? Glaciers ending within fiords are highly susceptible to slight changes in their immediate environment. Changes in water depth, water temperature, ice flux, and fiord geometry can cause profound responses (e.g. Meier *et al.* 1980; Brown *et al.* 1982; Mercer 1961). Denton and Hughes (1981), among others, have argued that ice shelves serve to buttress ice sheets and restrict the surging of ice streams. I suggest, if Figure 6 has merit, that fiord glaciers extended down-fiord during the period of ice shelf development. Indeed many of the fiord glaciers may have become ungrounded over the deep middle basins of the fiords (Løken and Hodgson 1971) as they pushed seaward protected by the "sea ice shelf". This process would cause buckling and pressure ridging of the fast ice within the fiord, in turn causing the ice to thicken.

Upon removal of the outer fast-ice sheet (Figure 7), fiord glaciers would be exposed to waves and winds and may have retreated catastrophically to points in the inner fiords where a combination of plan geometry and bathymetry enabled the glaciers to balance the losses due to calving with those associated with the transfer of mass from the ice sheet.

This scenario apparently applied in Frobisher Bay where radiocarbon dating suggests the ice front retreated down-Bay some 100 km in about 300 years - a net retreat of 330 m yr^{-1}.

One explanation for the drastic change between 10,700 and about 10,000 years ago is that there was a sudden and marked warming of the summer season. One site on the Clyde cliffs has a buried peaty soil with a pollen profile. This enables me to estimate July temperature based on the transfer functions discussed by Andrews *et al.* (1980). The resulting temperature estimate of 6.3°C (Figure 7) is distinctly *warmer* than the present

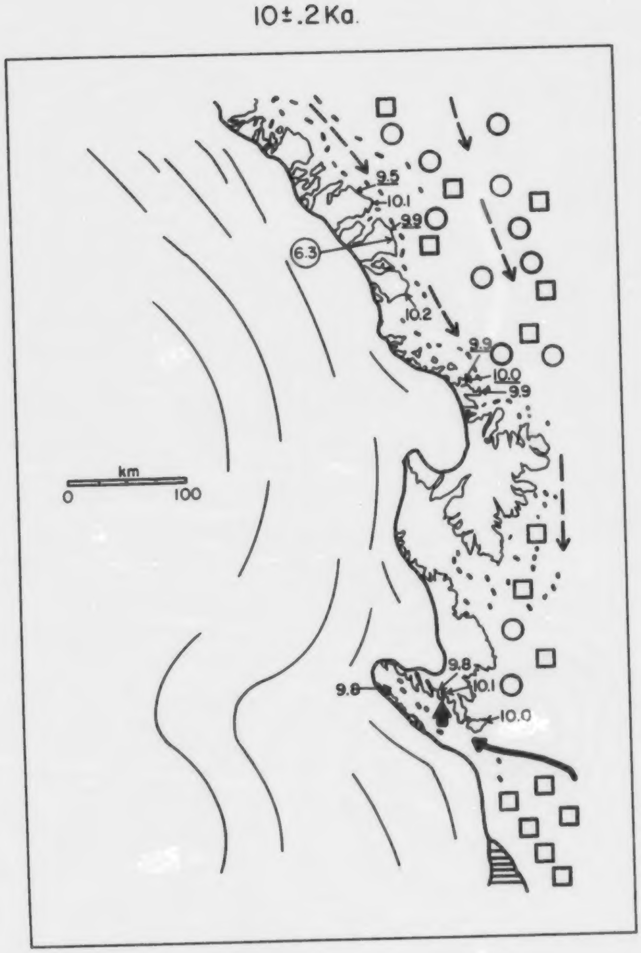

10±.2 Ka.
9.5
10.1
9.9
6.3
10.2
9.9
10.0
9.9
km
0
100
9.8
9.8
10.1
10.0

FIGURE 7: *The eastern Canadian Arctic about 10,000 years ago.*

July temperature. Mode (1980) also reported that pollen from an organic sample dated 9900 years ago on the Qivitu Foreland (Figure 7) contained more *Betula* and *Alnus* than present-day samples. These data also suggest that the climate had warmed appreciably some 10,000 years ago.

In the marine environment, Miller (1980) reported the first incursion of subarctic marine molluscs into outer Frobisher Bay about 9800 years ago. However, this warming is not seen farther north along the east coast until between 8600 and 8400 years ago (Andrews 1972; Miller 1980).

8000 YEARS AGO (COCKBURN SUBSTAGE, FIGURE 8)

Between approximately 9500 and 8000 years ago, evidently relative sea level was uniformly high within the fiords of Baffin Island (Andrews 1980). The latter part of the Cockburn substage (Figures 1, 8) was a period of major environmental change over much of the eastern Canadian Arctic. Evidence from several localities indicates that about 8000 years ago a series of glacial advances overran marine sediments (Smith 1966; Andrews *et al.* 1970B). These events can be traced in a series of massive end and lateral moraines that Ives and Andrews (1963) called the Cockburn moraines (see also: Falconer *et al.* 1965; Hodgson and Haselton 1974; Blake 1966; Miller and Dyke 1974; Miller 1980; Andrews *et al.* 1970B). These moraines are spectacular features that can be picked out on 1:1,000,000 LANDSAT imagery! Typically, the till ridges are 30-50 m high with widths of 100-200 m. Thus an average cross-section 1m-wide contains 3000 m^3 of till. Moraines of this size require significant time for their formation - especially those that lie on the inter-fiord plateau surface where ice flow would have been largely diverging. On these high uplands, the ice would have been thin and cold, thus flow would have been probably a few metres per year. Estimates of till deposition on Baffin Island (Barnett and Holdsworth 1974; Dowdeswell 1982) approach 10 $m^3 m^{-1} yr^{-1}$ at the base of glaciers within valleys. This suggests that the minimum time taken to form the vast Cockburn moraines is 300 years.

Paradoxically, at the same time as fiord glaciers were advancing in eastern Baffin Island it appears that the Laurentide ice sheet was in a terminal state of decline! Although ice from Hudson Strait did not recede from the extreme southeastern tip of Baffin Island until about 8300 years ago (Osterman *et al.* in press), by 8000 years ago the sea had penetrated Hudson Strait reaching the southern end of Hudson Bay, where, at this time, 8±.2 Ka.

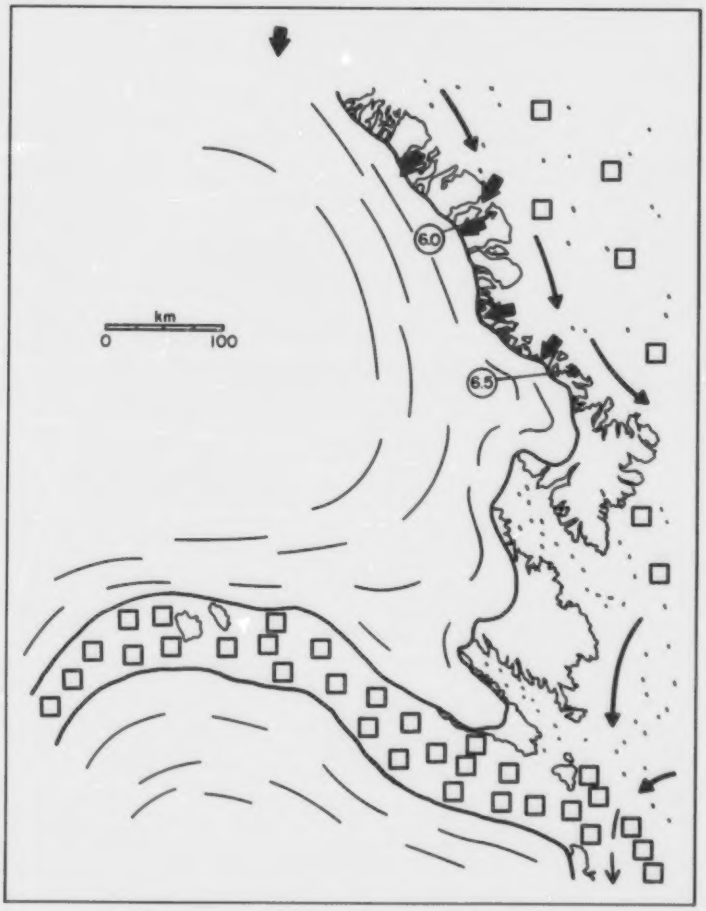

km
0
100
6.0
6.5

FIGURE 8: *The eastern Canadian Arctic about 8,000 years ago.*

the ice had been retreating northward, damming up the massive Glacial Lake Barlow-Ojibway (Vincent and Hardy 1979). However, radiocarbon dates of 8000 years ago have not been reported on the coasts of inner Hudson Strait. Thus, on Figure 8, I show the margin of the ice sheet still reaching tidewater along much of southern Baffin Island and northwestern Ungava. Ungava Bay was not deglaciated until 7500 years ago, whereas Foxe Basin still had ice until 6700 years ago.

Collections of molluscs from raised marine sediments along the eastern coast of Baffin Island indicate conclusively that, by 8400 to 8000 years ago, subarctic water was dominant along that coast (Figure 8). Species such as *Mytilus edulis, Chlamys islandicus, Macoma balthica,* and *Mya pseudoarenaria*, which no longer inhabit the waters off eastern and northern Baffin Island, appeared, and have been collected not only in distal marine strata but have been noted in foreset beds of ice-contact marine deltaic sequences. Evidently therefore, by late Cockburn time, the nearshore marine environment of eastern Baffin Island was like today (Andrews 1972; Andrews *et al.* 1981B) with sea surface temperatures warmer than present. This also strongly suggests that the extent of winter sea ice was less than today and that break-up occurred earlier than present.

These deductions are based on ecological requirements of molluscs that lived in 5-40 m of water. Logically, such ameliorated marine conditions would have had an impact on the terrestrial flora. Transfer-function estimates of July temperatures have been obtained for two sites (Figure 8), and indeed indicate that temperatures were 1-2°C warmer than present.

Why was glacier retreat in the Baffin Island fiords so relatively slow (10-20 m yr^{-1} during late Cockburn and throughout Kangilo time (Figure 1) if summer temperatures were warmer than present and if sea surface temperatures were higher? The conditions must surely have resulted in increased ablation on existing glaciers, many of which were also terminating in moderate water depths within fiords. Thus, calving would have probably been an additional important mechanism for mass loss. I (Andrews 1982B) suggest that the inference from slow deglaciation of the fiords must be that winter accumulation on the ice sheet was substantially higher than it had been during the latest Pleistocene.

6700 YEARS AGO (MIDDLE KANGILO SUBSTAGE, FIGURE 9)

The situation in the eastern Canadian Arctic 6700 years ago (Figure 9) shows a dramatic change in the degree of glacierization of the region. This interval occurs at about the

6.7±.2 Ka

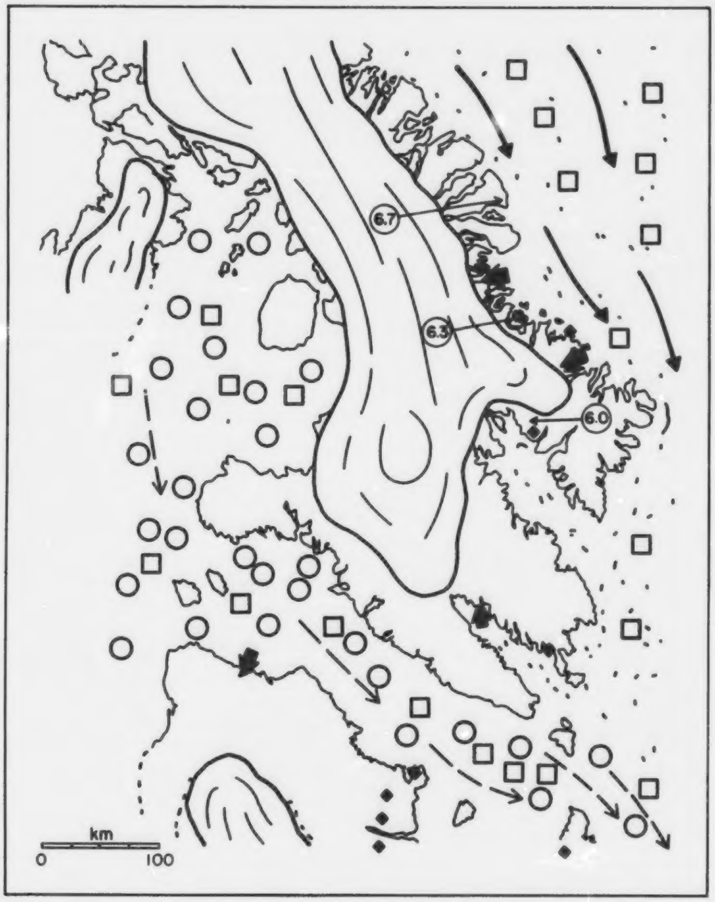

6.7
6.3
6.0
km
0
100

FIGURE 9: *The eastern Canadian Arctic about 6,700 years ago.*

middle of typical estimates for the Hypsithermal over much of the northern hemisphere. Rapid deglaciation of both coasts of Foxe Basin occurred about this time, and thus the Baffin Island ice sheet was restricted to an elongated mass running along the Island's axis.

Subarctic molluscs have been reported in collections from the eastern Baffin Island coast to Ungava (Figure 9). July temperature estimates, ranging from 6.0° to 6.7°C, suggest that summer conditions were as warm as today, if not warmer.

Of considerable interest to this scenario are the pollen reconstructions carried out along the Labrador coast and interior (Short 1978; Short and Nichols 1977; Stravers 1981). Macpherson (1982) has commented on this pollen evidence, specifically on the timing of the local climatic optimum, which was delayed with the period of maximum warmth occurring between 5300-3200 yr B.P. Macpherson (1982, P. 192) states the situation succinctly:

> "Thus, despite the general atmospheric warming between 8000 and 6500 B.P., or rather as a result of the warming, Labrador lay between a warming residual ice cap to the west and a cold ocean to the east."

Figures 8 and 9 suggest that one reason why the Labrador coast was so cold during this interval was because of the massive flux of icebergs, which must have drifted southward along the coast. Evidence from Baffin Island suggests, however, that marine climate was "warm" during this interval. Thus the flux of ice from the High Arctic was not of sufficient quantity to reduce the effects of a more expanded North Water/West Greenland Current. So, field data suggest that a "warm" Baffin Island lay north of a "cold" Labrador. However, by 6000 years ago, most ice had disappeared from the shores of Hudson Bay and much of Foxe Basin. This would have allowed the surface coastal waters off Labrador to have warmed considerably.

3000 YEARS AGO (IGLUTALIK SUBSTAGE, FIGURE 10)

The final "snapshot" of the eastern Canadian Arctic is taken prior to the major deterioration in climate that occurred between 2500 and 2000 years ago (Andrews *et al.* 1981A). By 3000 years ago, the Laurentide ice sheet had shrunk to one or two remnant ice masses, including the Barnes and Penny ice caps. However, local valley and cirque glaciers had expanded before 3200 years ago (Miller 1973), and thereafter fluctuated several times throughout the neoglaciation (Andrews 1982B). July temperature estimates (Figure 10) and pollen data (Davis 1980; Mode 1980) suggest that, by 3000 years ago, vegetation showed

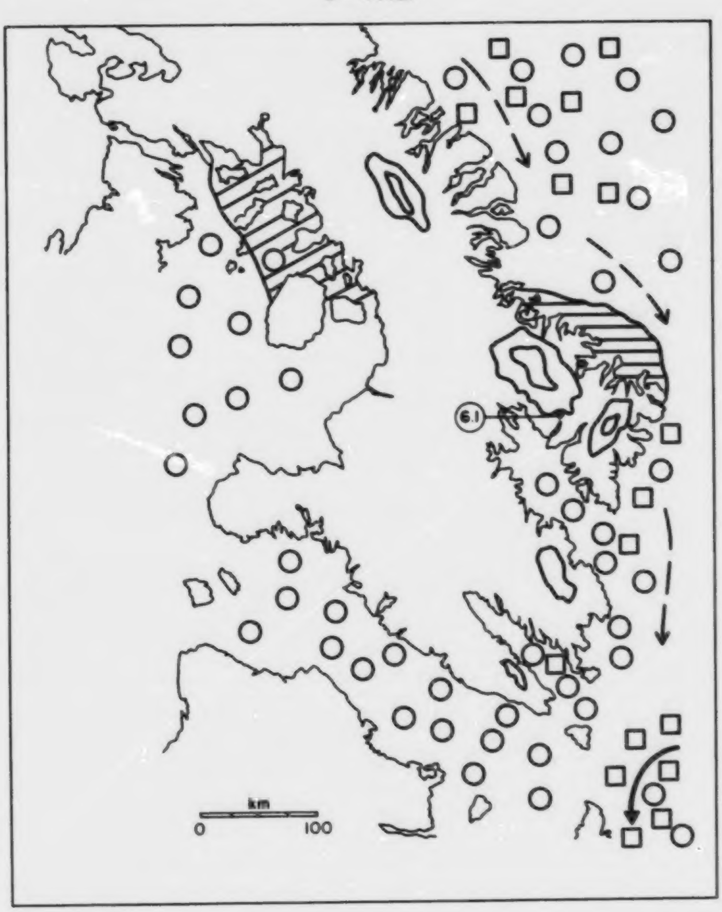

3±.1Ka.
6.1
km
0
100

FIGURE 10: *The eastern Canadian Arctic about 3,000 years ago.*

little evidence for a great climatic deterioration, whereas between 2500 and 2000 years ago significant changes are apparent in pollen diagrams (Short and Andrews 1981; Short and Jacobs 1982; Davis 1980). In addition, extensive spreads of niveo-eolian sands suggest that during the Windy Lake substage climate became colder and drier.

CONCLUSIONS

Maps of the paleogeography of the eastern Canadian Arctic that I have presented are based on a variety of data. They suggest that there have been major environmental changes over the last 11,000 years. Although the history of glacial advances and retreats is relatively well known for the last 8000 years, we have much to learn about the events prior to that time. Continued field research will shed some light on that interval; probably, however, the most important insights will emerge after detailed study of continuous lake and marine cores. These studies are now underway at various institutions in Canada and the United States.

ACKNOWLEDGEMENTS

My research in the eastern Canadian Arctic has been largely sponsored by the United States National Science Foundation. This paper is a contribution to: ATM-82-08677, DPP-81-116048 and EAR-81-21296. I thank my colleagues at INSTAAR (Institute of Arctic and Alpine Research, Boulder, Colorado) for the many discussions held on these and other topics.

REFERENCES

Aksu, A. 1981. Late Quaternary stratigraphy, paleoenvironmentology and sedimentation history of Baffin Bay and Davis Strait. Ph.D. thesis, Dalhousie University, Halifax. 771 pp.

Andrews, J.T. 1970. A geomorphological study of post-glacial uplift, with particular reference to arctic Canada. Institute of British Geographers, Special Publication No. 2:1-156.

______. 1972. Recent and fossil growth rates of marine bivalves, Canadian Arctic, and Late Quaternary arctic marine environments. Palaeogeography, Palaeoclimatology, Palaeoecology 11:157-176.

______. 1975. Support for a stable late Wisconsin ice margin (14,000 to ca 9,000 BP): a test based on glacial rebound. Geology 4:617-620.

Andrews, J.T. 1980. Progress in relative sea level and ice sheet reconstructions, Baffin Island, N.W.T., for the last 125,000 years. In: Earth Rheology, Isostasy and Eustasy. Edited by: N.-A. M rner. John Wiley, New York. pp. 175-200.

_______. 1982A. Chronostratigraphic division of the Holocene, arctic Canada. Striae 16:56-64.

_______. 1982B. Holocene glacier variations in the eastern Canadian Arctic: a review. Striae 18:9-14.

Andrews, J.T., and D.M. Barnett. 1979. Holocene (neoglacial) moraine and proglacial lake chronology, Barnes Ice Cap, Canada. Boreas 8:339-356.

Andrews, J.T., R.G. Barry, and L. Drapier. 1970. An inventory of the present and past glacierization of Home Bay and Okoa Bay, east Baffin Island, N.W.T., Canada, and some climatic and paleoclimatic considerations. Journal of Glaciology 9:337-362.

Andrews, J.T., J.T. Buckley, and J.H. England. 1970. Late-glacial chronology and glacio-isostatic recovery, Home Bay, east Baffin Island. Geological Society of America Bulletin 81:1123-1148.

Andrews, J.T., P.T. Davis, W.N. Mode, H. Nichols, and S.K. Short. 1981A. Relative departures of July temperature in northern Canada for the past 6,000 yr. Nature 289:164-167.

Andrews, J.T., G.H. Miller, A.R. Nelson, W.N. Mode, and W.W. Locke, III. 1981B. Quaternary near-shore environments on eastern Baffin Island, N.W.T. In: Quaternary Paleoclimate. Edited by: W.C. Mahaney. Geoabstracts, Norwich, England. pp. 13-44.

Andrews J.T., P.T. Davis, and C. Wright. 1976. Little Ice Age permanent snow cover in the eastern Canadian Arctic: extent mapped from LANDSAT-1 satellite imagery. Geografiska Annaler 58:71-81.

Andrews J.T., and G.H. Miller. 1972. Maps of the present glaciation limits and lowest equilibrium line altitude for north and south Baffin Island. Arctic and Alpine Research 4:45-60.

Andrews, J.T., W.N. Mode, and P.T. Davis. 1980. Holocene climate based on pollen transfer function equations, eastern Canadian Arctic. Arctic and Alpine Research 12:41-64.

Andrews, J.T., and S.K. Short. (In press). Radiocarbon date list V from Baffin Island; radiocarbon date list II from Labrador. Institute of Arctic and Alpine Research, Occasional Paper, University of Colorado, Boulder.

Barnett, D.M., and G. Holdsworth. 1974. Origin, morphology, and chronology of sublacustrine moraines, Generator Lake, Baffin Island, Northwest Territories, Canada. Canadian Journal Earth Sciences 11:380-408.

Blake, W., Jr. 1966. End moraines and deglaciation chronology in northern Canada with special reference to southern Baffin Island. Geological Survey of Canada Paper 66-26:1-31.

______. 1970. Studies of glacial history of arctic Canada. I. Pumice, radiocarbon dates, and differential postglacial uplift in the eastern Queen Elizabeth Islands. Canadian Journal Earth Sciences 7:634-664.

______. 1975. Radiocarbon age determinations and postglacial emergence at Cape Storm, southern Ellesmere Island. Geografiska Annaler 62A:1-71.

Brown, C.S., M.F. Meier, and A. Post. 1982. Calving speed of Alaska tidewater glaciers, with application to Columbia Glacier. United States Geological Survey Professional Paper 1258-C:1-13.

Cline, R.M., and J.E. Hays. (Editors). 1976. Investigation of Late Quaternary paleoceanography and paleoclimatology. Geological Society of America Memoir 145:1-464.

Craig, B.G., and J.G. Fyles. 1960. Pleistocene geology of arctic Canada. Geological Survey of Canada Paper 60-10:1-21.

Crane, R.G. 1978. Seasonal variations in sea ice extent in the Davis Strait-Labrador Sea area and relationships with synoptic-scale atmospheric circulation. Arctic 31(4):434-447.

Crary, A.P. 1960. Arctic ice island and ice shelf studies. Part II. Arctic 13:32-50.

Davis, P.T. 1980. Late Holocene glacial, vegetational, and climatic history of Pangnirtung and Kingnait fiord area, Baffin Island, N.W.T., Canada. Ph.D. thesis, University of Colorado, Boulder. 366 pp.

Denton, G.H., and T.J. Hughes. (Editors). 1981. The last great ice sheets. John Wiley, New York. 484 pp.

Dowdeswell, J.A. 1982. Debris transport paths and sediment flux through the Grinnell Ice Cap, Frobisher Bay, Baffin Island, N.W.T., Canada. M.A. thesis, University of Colorado, Boulder. 176 pp.

Dyke, A.S., J.T. Andrews, and G.H. Miller. 1982. Quaternary geology of Cumberland Peninsula, Baffin Island, District of Franklin. Geological Survey of Canada Memoir 403:1-32.

England, J.H. 1976. Late Quaternary glaciation of the eastern Queen Elizabeth Islands, N.W.T., Canada: alternative models. Quaternary Research 6:185-203.

______. 1983. Isostatic adjustments in a full glacial sea. Canadian Journal of Earth Sciences 20:895-917.

Falconer, G. 1962. Glaciers of northern Baffin and Bylot Islands, N.W.T. Canadian Geographical Branch (Ottawa) Paper 33:1-31.

Falconer, G., J.D. Ives, O.H. Løken, and J.T. Andrews. 1965. Major end moraines in eastern and central arctic Canada. Geographical Bulletin 7:137-153.

Fillon, R.H., and J.C. Duplessy 1980. Labrador Sea bio-, tephro-, oxygen isotope stratigraphy and late Quaternary paleoceanographic trends. Canadian Journal of Earth Sciences 17:831-854.

Fillon, R.H., and R.A. Harmes. 1982. Northern Labrador shelf glacial chronology and depositional environments. Canadian Journal of Earth Sciences 19:162-192.

Gilbert, R. 1982. Contemporary sedimentary environments on Baffin Island, N.W.T., Canada; glaciomarine processes in fiords of eastern Cumberland Peninsula. Arctic and Alpine Research 14:1-12.

Harrison, D.A. 1964. A reconnaissance glacier and geomorphological survey of the Duart Lake area, Bruce Mountains, Baffin Island, N.W.T. Geographical Bulletin No. 22:57-71.

Hodgson, D.A., and G.M. Haselton. 1974. Reconnaissance glacial geology, northeastern Baffin Island. Geological Survey of Canada Paper 74:20:1-10.

Ives, J.D. 1962. Indications of recent extensive glacierization in north-central Baffin Island, N.W.T. Journal of Glaciology 4:197-205.

Ives, J.D., and J.T. Andrews. 1963. Studies in the physical geography of north-central Baffin Island. Geographical Bulletin 19:5-48.

Ives, J.D., J.T. Andrews, and R.G. Barry. 1975. Growth and decay of the Laurentide ice sheet and comparisons with Fennoscandia. Naturwissenschaften 62:118-125.

Jacobs, J.D., and J.P. Newell. 1979. Recent year-to-year variations in seasonal temperatures and sea ice conditions in the eastern Canadian Arctic. Arctic 32(4):345-354.

Koerner, R.M., and D.A. Fisher. 1981. Studying climatic change from Canadian High Arctic ice cores. In: Climatic Change in Canada 2. Edited by: C.R. Harington. Syllogeus No. 33:195-218.

Locke, C., and W.W. Locke, III. 1977. Little Ice Age snow-cover extent and paleoglacierization thresholds: north-central Baffin Island, N.W.T., Canada. Arctic and Alpine Research 9:291-300.

Locke, W.W., III. 1980. The Quaternary geology of the Cape Dyer area, southeasternmost Baffin Island, Canada. Ph.D. thesis, University of Colorado, Boulder. 331 pp.

Locke, W.W., III, J.T. Andrews, and P.J. Webber. 1979. A manual for lichenometry. GeoAbstracts, Norwich, England. Technical Bulletin No. 26:1-47.

Løken, O.H., and D.A. Hodgson. 1971. On the submarine geomorphology along the east coast of Baffin Island. Canadian Journal of Earth Sciences 8:185-195.

Lyons, J.B., and J.E. Mielke. 1973. Holocene history of a portion of northernmost Ellesmere Island. Arctic 26:314-323.

Macpherson, J. Brown. 1982. Postglacial vegetational history of the eastern Avalon Peninsula, Newfoundland, and Holocene climatic change along the eastern Canadian seaboard. Géographie physique et Quaternaire 36:175-196.

Markham, W.E. 1981. Ice atlas, Canadian Arctic waterways. Canadian Government Publishing Centre, Hull. 198 pp.

Meier, M.F., *et al.* 1980. Predicted timing of the disintegration of the lower reach of Columbia Glacier, Alaska. United States Geological Survey, Open File Report 80-582:1-47.

Mercer, J.H. 1956. Geomorphology and glacial history of southernmost Baffin Island. Geological Society of America Bulletin 67:553-570.

______. 1961. The response of fjord glaciers to changes in the firn limit. Journal of Glaciology 3:850-866.

______. 1983. Cenozoic glaciation in the southern hemisphere. Annual Review Earth and Planetary Sciences 11:99-132.

Miller, G.H. 1973. Late Quaternary glacial and climatic history of northern Cumberland Peninsula, Baffin Island, N.W.T., Canada. Quaternary Research 3:561-583.

______. 1980. Late Foxe glaciation of southern Baffin Island, N.W.T., Canada. Geological Society of America Bulletin, Part I, 91:399-405.

Miller, G.H., and A.S. Dyke. 1974. Proposed extent of late Wisconsin Laurentide ice on eastern Baffin Island. Geology 2:125-130.

Mode, W.N. 1980. Quaternary stratigraphy and palynology of the Clyde Foreland, Baffin Island, N.W.T., Canada. Ph.D. thesis, University of Colorado, Boulder. 219 pp.

Nelson, A.R. 1982. Aminostratigraphy of Quaternary marine and glaciomarine sediments, Qivitu Peninsula, Baffin Island. Canadian Journal of Earth Sciences 19:945-961.

Nichols, H., P.M. Kelly, and J.T. Andrews. 1978. Holocene palaeowind evidence from palynology. Nature 273:140-142.

Osterman, L.E. 1982. Late Quaternary history of southern Baffin Island, Canada: a study of foraminifera and sediments from Frobisher Bay. Ph. D. thesis, University of Colorado, Boulder. 380 pp.

Osterman, L.E., G.H. Miller, and J.A. Stravers. (In press). Middle and late Foxe glacial events in southern Baffin Island. In: Quaternary Studies in Baffin Island, West Greenland, and in Baffin Bay. Edited by: J.T. Andrews and M.L. Andrews.

Peltier, W.R., and J.T. Andrews. 1976. Glacial isostatic adjustment. I. The forward problem. Geophysical Journal of the Royal Astronomical Society 46:605-646.

Quinlan, G.M. 1981. Numerical models of postglacial relative sea level change in Atlantic Canada and the eastern Canadian Arctic. Ph.D. thesis, Dalhousie University, Halifax. 499 pp.

Quinlan, G., and C. Beaumont. 1982. The deglaciation of Atlantic Canada as reconstructed from the postglacial relative sea-level record. Canadian Journal of Earth Sciences 19:2232-2248.

Rothlisberger, H. 1951. Quatärgeologische beobachtungen am Eglinton Fiord, Baffin-Land, Kanada. Naturforschende Gesellschaft in Zurich, Vierteljahrsschrift 96:252-257.

Short, S.K. 1978. Palynology: a Holocene environmental perspective for archeology in Labrador-Ungava. Arctic Anthropology 15:9-35.

Short, S.K., and J.T. Andrews. 1980. Palynology of six middle and late Holocene peat sections, Baffin Island. Géographie physique et Quaternaire 34:61-75.

Short, S.K., and J.D. Jacobs. 1982. A 1100-year paleoclimatic record from Burton Bay-Tarr Inlet, Baffin Island. Canadian Journal of Earth Sciences 19:398-409.

Short, S.K., and H. Nichols. 1977. Holocene pollen diagrams from subarctic Labrador-Ungava: vegetational history and climatic change. Arctic and Alpine Research 9:265-290.

Smith, J.E. 1966. Sam Ford Fiord: a study in deglaciation. M.Sc. thesis, McGill University, Montréal. 93 pp.

Stravers, J.A., and G.H. Miller. 1983. Late Foxe/Wisconsin glacial ice masses of outer Meta Incognita Peninsula and adjacent Frobisher Bay and Hudson Strait, southern Baffin Island, eastern Canadian Arctic. Geological Society of America, Abstracts with Programs 15(6):699.

Stravers, L.K.S. 1981. Palynology and deglaciation history of central Labrador-Ungava Peninsula. M.Sc. thesis, University of Colorado, Boulder. 171 pp.

Stuckenrath, R., G.H. Miller, and J.T. Andrews. 1979. Problems of radiocarbon dating Holocene organic-poor sediments, Cumberland Peninsula. Arctic and Alpine Research 11:109-120.

Stuiver, M., G.H. Denton, T.J. Hughes, and J.L. Fastook. 1981. History of the marine ice sheet in west Antarctica during the last glaciation: a working hypothesis. In: The Last Great Ice Sheets. Edited by: G.H. Denton and T.J. Hughes. John Wiley, New York. pp. 319-439.

Vincent, J.-S., and L. Hardy. 1979. The evolution of Glacial Lake Barlow and Ojibway, Quebec and Ontario. Geological Survey of Canada Bulletin 136:1-18.

Webb, T., III, and R.A. Bryson. 1972. Late and postglacial change in the northern Midwest, U.S.A.: quantitative estimates derived from fossil pollen spectra by multivariate statistical analysis. Quaternary Research 2:70-115.

Williams, L.D. 1978. The Little Ice Age glaciation level on Baffin Island, arctic Canada. Palaeogeography, Palaeoclimatology, Palaeoecology 25:199-207.

OLD CROW TEPHRA: ITS SIGNIFICANCE FOR UNDERSTANDING THE EARLY AND MIDDLE WISCONSINAN CLIMATE HISTORY OF EASTERN BERINGIA

John V. Matthews, Jr.[1] and Charles E. Schweger[2]

INTRODUCTION

Certain types of geological phenomena allow one to delineate specific "instants" in geological time. Geomagnetic polarity reversals and volcanic tephras are two well known examples. One volcanic tephra that is gaining importance for understanding the late Quaternary climatic history of eastern Beringia (Alaska and the Yukon) is the Old Crow tephra. It occurs at a number of sites in eastern Beringia, and was apparently deposited during the early Wisconsinan - a period for which our resolution of climatic events is especially blurry. In addition to its value for describing climate and vegetation at the instant of the ash fall, Old Crow tephra contributes to a knowledge of climatic events that occurred long after its deposition. Also, because at several localities the tephra is associated with glacial drift, it provides new information on the timing of glacial events in eastern Beringia.

OLD CROW TEPHRA - DISTRIBUTION AND AGE

Old Crow tephra was first discovered by O.L. Hughes (Geological Survey of Canada, Calgary) during a study of the HH 228 ("Big Bluff") exposure on the Porcupine River in the northern Yukon Territory (Figure 1). It was later found at several exposures within the Old Crow Basin north of the HH 228 site, and recently tephra collected originally by V.N. Rampton in the Kluane Lake region of the southern Yukon has also been identified as the Old Crow type (Westgate 1982). In Alaska, Old Crow tephra occurs at exposures in the Koyukuk River valley (Westgate *et al.* 1983), in the Fairbanks region (Westgate *et al.* 1982), and in the lower part of core V at Imuruk Lake (Figure 1) (Shackleton 1982; Westgate 1982).

The tephra is beyond the range of radiocarbon dating, but dates stratigraphically above it show that it was deposited more than 60,000 years ago (Westgate *et al.* 1983), whereas a

---

[1] Terrain Sciences Division, Geological Survey of Canada, Ottawa, Ontario K1A 0E8

[2] Department of Anthropology, University of Alberta, Edmonton, Alberta T6G 2H4

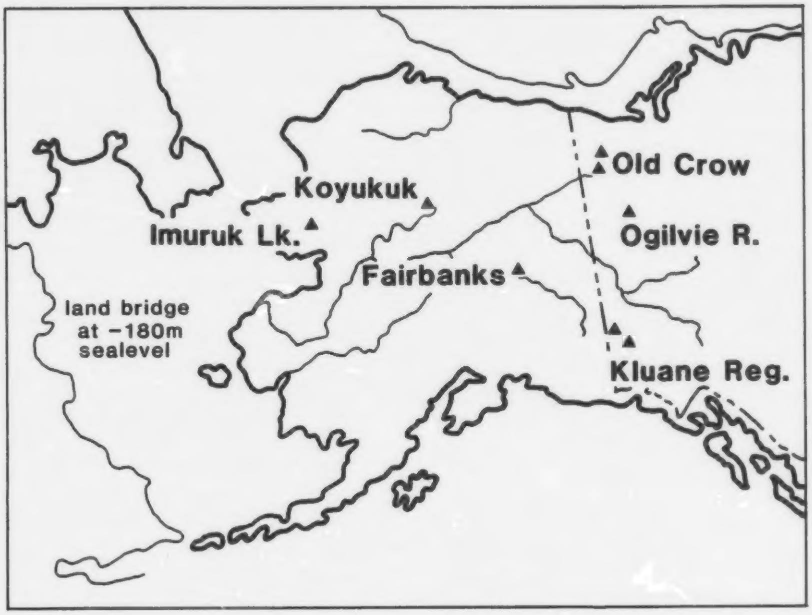

Old Crow
Koyukuk
Imuruk Lk.
Ogilvie R.
Fairbanks
land bridge
at -180m
sealevel
Kluane Reg.

FIGURE 1: *Distribution of Old Crow Tephra in eastern Beringia (Westgate 1982). The HH 228, or "Big Bluff" section, is represented by the most southern of the two locality markers in the Old Crow region.*

fission-track date on the tephra itself reveals it to be very likely less than 120,000 years old (Naeser *et al.* 1982). More precise estimates of its age exist (e.g., 70,000 and 80,000 yr B.P., cited in Westgate 1982), but they are little more than educated guesses. A more defensible age estimate has been obtained by interpolating radiocarbon and paleomagnetically-dated levels in the Imuruk core V (Schweger and Matthews, Ms.). It suggests that the volcanic eruption (or closely-spaced eruptions) responsible for the tephra occurred some time between 87,000 and 105,000 years ago, an interval that corresponds to the latter part of marine isotopic Stage 5 (Figure 2).

PALEOECOLOGICAL IMPLICATIONS

Pollen and macrofossil evidence from the Old Crow region and pollen spectra from one of the Koyukuk valley sections and Imuruk core V show that, at the "instant" of the Old Crow tephra deposition, those three areas (Figure 1) were vegetated with shrub tundra containing a significant herb component. Two of the regions (Old Crow and Koyukuk) are presently forested, so former tundra conditions imply colder climate. That the climate was probably colder than even currently low arctic tundra, is indicated by the rarity of alder pollen. As yet closely-associated climate proxy data is not available at the more southern occurrence of the tephra. When such information is in hand, it should be possible to provide a reasonable approximation of synoptic climate at the "instant" of the ash fall.

Many of the exposures containing Old Crow tephra display a rich and varied paleoclimatic record. Those in the Old Crow region show that deposition of the tephra was followed (probably thousands of years later) by a period of climatic warming marked by permafrost degradation and slight northward displacment of floral and faunal elements (Matthews 1980). Deep thawing and slumping also occurred in the Fairbanks region after deposition of Old Crow tephra, but prior to 56,900 years (Schweger and Matthews, Ms.). The pollen sequence above the tephra at Koyukuk section KY-11 reveals an extended interval of cold climate punctuated by several minor pulses of spruce. It ends about 60,000 yr B.P. with an abrupt rise of spruce and alder percentages to levels comparable to those of the present (Schweger and Matthews, Ms.). The Imuruk Lake sequence is similar. Zone il, with high spruce and alder percentages, follows a lengthy tundra interval that has the Old Crow tephra near its base. In this case, however, spruce and alder percentages are considerably higher than in local

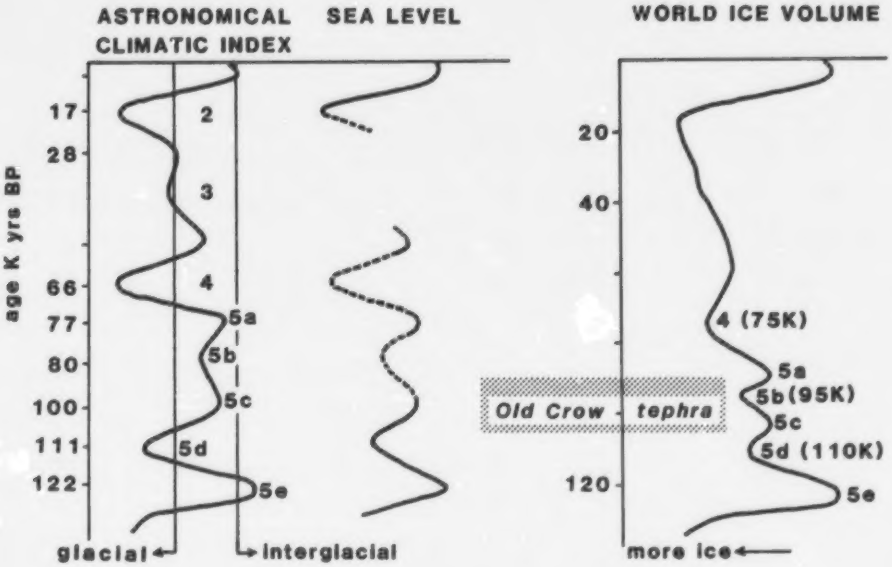


FIGURE 2: *Comparison of marine isotopic stages (after Dansgaard and Duplessy 1981) and the Astronomical Climatic Index (after Kukla 1979) with the extrapolated age for the Old Crow Tephra (shaded). Note that the vertical scales are not identical and that there are some important areas of disagreement between the two reconstructions. Both imply, however, that the last period of world climate as warm as at present occurred during Stage 5e — before deposition of the Old Crow Tephra.*

surface samples, signifying a warmer climate than at present. The extrapolated age of zone il is approximately 60,000 yr B.P. (Schweger and Matthews, Ms.).

All of this evidence of warming appears to be a manifestation of a single event. Every one of the fluctuations occurs above the Old Crow tephra, and each one represents the same type of climatic change. We view it as a single pulse of warm climate, sensed as an interglacial in some areas, that occurred about 60,000 years ago in eastern Beringia and possibly other parts of the northern hemisphere (Dreimanis and Raukas 1975). The date of 60,000 years places it at the start of marine isotope Stage 3, an interval that, in terms of world ice-volume, does not appear as warm as we suggest. This discrepancy cannot be adequately explained now, but it may be due to peculiarities of the geography of the Beringian region (especially the land bridge) during mid-Wisconsinan time (Schweger and Matthews, Ms.).

In the southern Yukon, Old Crow tephra occurs stratigraphically above drift of the Mirror Creek glaciation (Rampton 1971), a local glaciation which is thought to correlate with the Reid and Buckland glaciations in other parts of the Yukon, and with the Delta Glaciation in the central Alaska Range and the Itkillik Glaciation in the Brooks Range (Hamilton 1982). All of these glaciations are beyond the limit of radiocarbon dating, and have been assigned to either the early Wisconsinan or the Illinoian. If our conclusion concerning the age of the Old Crow tephra is correct, these various local glaciations occurred during marine Stage 5 and not during Stage 4 as might be assumed by inspection of the isotopic ice-volume curve (Figure 2).

CONCLUSION

Old Crow tephra has become an important marker horizon in eastern Beringia. It has both direct and indirect paleoclimatic and paleoecological implications. Like many other tephras, it does not coincide with a particular climatic event or threshold; and consequently, most workers would probably not rank the "instant" of the tephra fall as one of the "critical periods" to be discussed in this symposium. We submit that any period that is as nearly instantaneous as that represented by a tephra and as widely recognized in the geological record is a "critical period", because it has the potential for providing a unique insight into past climatic conditions. For example, the Old Crow tephra has the potential for revealing synoptic climate at an instant within a time interval that is poorly documented by other evidence. And as indicated above, Old Crow tephra provides new insight into climatic events postdating and predating it.

ACKNOWLEDGEMENTS

This work has benefited from the generous attitude of many colleagues who freely shared their data, results, and opinions. We express special thanks to Paul Colinvaux, Thomas Hamilton, Owen Hughes, Sigrid Lichti-Federovich, Richard Morlan, John Westgate, and Vern Rampton.

REFERENCES

Dansgaard, W., and J.-C. Duplessy. 1981. The Eemian interglacial and its termination. Boreas 10:219-228.

Dreimanis, A., and A. Raukas. 1975. Did middle Wisconsin, middle Weichselian, and their equivalents represent an interglacial or an interstadial complex in the northern hemisphere? In: Quaternary Studies. Edited by: R.P. Suggate and M.M. Cresswell. The Royal Society of New Zealand Bulletin 13:109-120.

Hamilton, T.D. 1982. A late Pleistocene glacial chronology for the southern Brooks Range: stratigraphic record and regional significance. Geological Society of America 93: 700-716.

Kukla, G. 1979. Probability of expected climatic stresses in North America in the next one M.Y. In: A Summary of FY-1978 Consultant Input for Scenario Methodology Development. Edited by: B.L. Scott, R.A. Craig, G.L. Benson and M.A. Harwell. Pacific Northwest Laboratory of Batelle Memorial Institute PNL-2851, VC-70:XIII:1-8.

Matthews, J.V., Jr. 1980. An early Wisconsinan warm interstadial in East Beringia? VI Biennial Meeting of the American Quaternary Association (AMQUA), Orono, Maine. Abstracts:130-131.

Naeser, N.D., J.A. Westgate, and O.L. Hughes. 1982. Fission-track ages of late Cenozoic distal tephra beds in the Yukon Territory and Alaska. Canadian Journal of Earth Sciences 19:2167-2178.

Rampton, V.N. 1971. Late Pleistocene glaciations of the Snag-Klutlan area, Yukon Territory. Arctic 24:277-300.

Schweger, C.E., and J.V. Matthews, Jr. (Ms.) Old Crow tephra and its implications for the synoptic paleoecology/paleoclimatology of Alaska and the Yukon during mid-Wisconsinan time.

Shackleton, J. 1982. Environmental histories from Whitefish and Imuruk Lakes, Seward Peninsula, Alaska. Ohio State University, Institute of Polar Studies Report No. 76: 1-49.

Westgate, J.A. 1982. Discovery of a large-magnitude, late Pleistocene volcanic eruption in Alaska. Science 218:789-790.

Westgate, J.A., T.D. Hamilton, and M.P. Gorton. 1983. Old Crow tephra: a new late Pleistocene stratigraphic marker across north-central Alaska and western Yukon Territory. Quaternary Research 19:38-54.

Westgate, J.A., T.L. Péwé, and M.P. Gorton. 1982. Tephrochronology of the Gold Hill Loess in central Alaska. Geological Society of America, 95th Annual Meeting, Louisiana. Abstracts: 645-646.

PALEOCLIMATOLOGY AND GLACIOLOGY

1550-1620: A PERIOD OF SUMMER ACCUMULATION IN THE QUEEN ELIZABETH ISLANDS

Bea Taylor Alt[1]

INTRODUCTION

The Little Ice Age interval is generally considered to comprise the period 1550-1850. Time series of the O^{18}/O^{16} ratio and melt characteristics measured in ice cores taken from Canadian High Arctic ice caps (Koerner 1977; Koerner and Fisher 1981; Paterson *et al.* 1977) provide an opportunity to investigate the climate of the Little Ice Age in the eastern Queen Elizabeth Islands (Figure 1).

δO^{18}

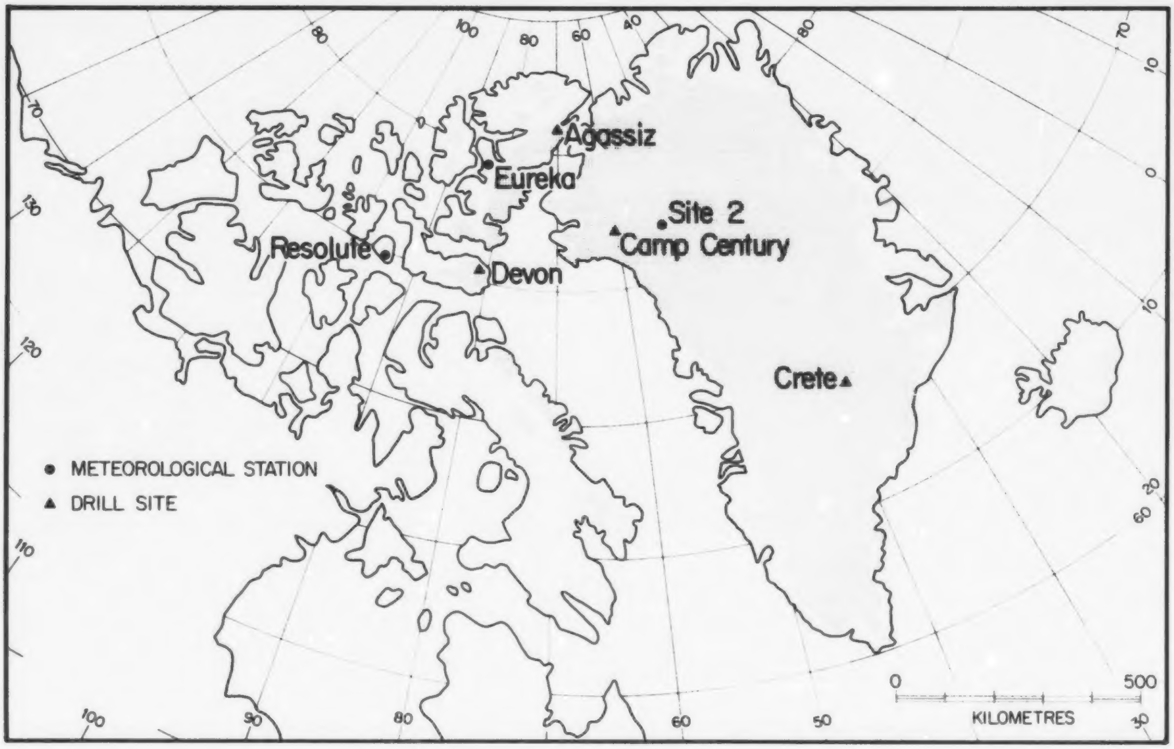
The O^{18}/O^{16} ratio is expressed as the fractional difference between the ratio in the sample and the ratio of 'standard mean ocean water'. It is referred to as δ and measured in $^{o}/oo$. In polar snow δ is negative; the more negative, the lower the temperature. The δ value in snow which falls on the ice cap is dependent on many factors, as is noted in all presentations of ice-core data (e.g. Dansgaard *et al.* 1973; Paterson *et al.* 1977; Fisher and Koerner 1983). Those which are relevant for the present discussion are:

(a) the temperature of condensation;
(b) the distance water has travelled (farther travelled - more negative δ); and
(c) the source area of precipitation.

In spite of the complexity of these and the geophysical factors, δ time series have been related to records of mean annual temperature with some success (Dansgaard *et al.* 1973).

---

[1] Polar Continental Shelf Project, 880 Wellington Street, Ottawa, Ontario K1A 0E4



Agassiz
Eureka
Site 2
Camp Century
Resolute
Devon
Crete
● METEOROLOGICAL STATION
▲ DRILL SITE
0
500
KILOMETRES
100
80
60
40
80
70
10
0
130
120
110
100
90
80
60
50
30
20
60
70

FIGURE 1: *Location map.*

PERCENT MELT (PC)

Ice formed by melting and refreezing can be distinguished from that formed by compaction of snow into ice. The thickness of such melt layers in the core has been measured by Koerner (1977), and is expressed as a percent of the annual accumulation layer thickness. PC (percent melt) is related to melt-season temperature or melting degree-days.

The PC time series from Devon Island Ice Cap (Figures 2A, B) shows three distinct minima (i.e. summers with little melt) within the period usually ascribed to the Little Ice Age. On the δ curve, the two most recent minima are evident, but during the older PC minimum, about 1550-1620, the δs remained variable, being more often above the 800-year average than below. This inverse relationship of δ and PC shows also in the 50-year average curves (Figures 2C, D).

The Agassiz Ice Cap core shows a similar relationship between δ and PC (Fisher and Koerner 1983, Figures 3, 4) during the 1550-1620 period. The Camp Century (Fisher 1982, Figure 2, p. 423) and Crete (Dansgaard *et al.* 1975, p. 25, Figure 2) ice cores from Greenland, on the contrary, do not exhibit this effect.

In this paper I examine the period 1550-1620, not only because of its importance as a preliminary phase of the Little Ice Age, but also to illustrate how ice-core values can be used to give climatic information - despite problems with transfer functions. The following discussions are based on data from Devon Island Ice Cap.

CLIMATE COMBINATIONS PRODUCING LESS NEGATIVE (WARM) δ AND LOW (COLD) PC VALUES

The following considerations point to climatic combinations which are consistent with warm δ values and little summer melt:

(1) Low melt indicates cold and/or short summers.

(2) If proportionally more solid precipitation (snow) falls in the warm season, the annual δ values will be weighted towards warm temperatures.

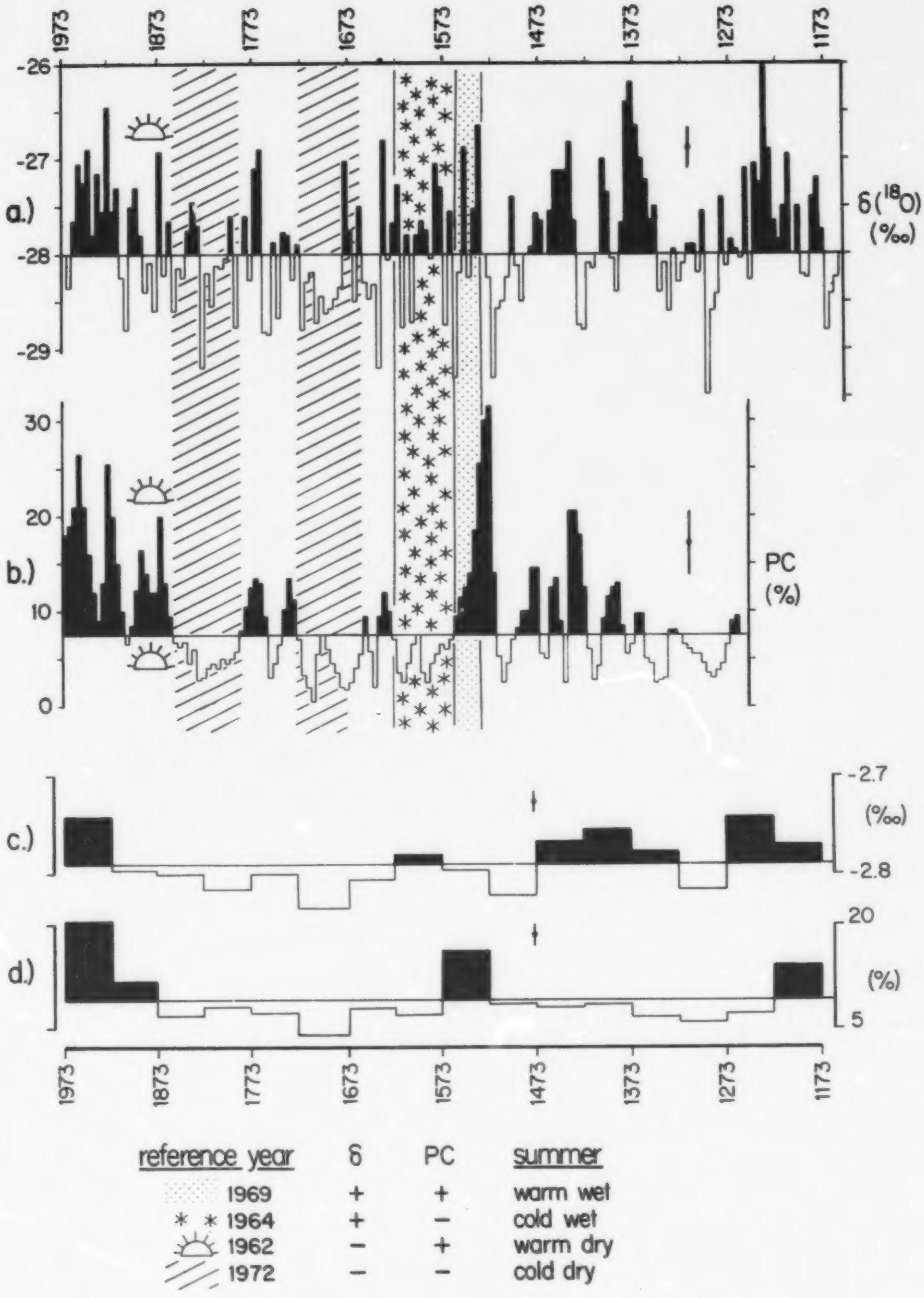
(3) If relatively more of that precipitation is from a local source, the δ values will be warmer (i.e. it is not as depleted of O^{18} as distant moisture).

(4) Very warm winters could also produce relatively warm δ values.

As shown in the Camp Century and Crete cores mentioned earlier, warm δ values between 1550-1620 are confined to the Queen Elizabeth Islands, which is consistent with a local effect (factor 3). Factors (1) through (3) suggest a cold, wet summer. This could be

FIGURE 2: *Devon ice core time series. The time scale is discussed in Paterson* et al. *(1977) and is accurate within 10%.*

a) combined δ records from 1972 and 1973 cores.
b) summer melt percent (PC) from 1973 core.
a) and b) are five-year averages after Koerner and Fisher (1981). Error bars represent the error in a single given value due to stratigraphic features.
c) combined δ record from 1972 and 1973 cores.
d) combined PC from 1971, 1972 and 1973 cores.
c) and d) are 50-year averages after Fisher (1977). The error in the 50-year values due to stratigraphic features is considerably smaller as indicated by the error bars.



1973
1873
1773
1673
1573
1473
1373
1273
1173
-26
-27
-28
-29
a.)
δ(^{18}O)
(‰)
30
20
10
0
b.)
PC
(%)
c.)
-2.7
(‰)
-2.8
d.)
20
(%)
5
reference year
δ
PC
summer
1969 + + warm wet
1964 + − cold wet
1962 − + warm dry
1972 − − cold dry

combined with a normal or warm winter. The only possible combination that does not include a wet summer would be a very warm winter followed by a cold dry summer. Meteorologically, this is an unlikely combination. Examination of the Resolute Bay temperature record (32 years) shows that the only winter with a temperature averaging more than 1°C above normal, which was followed by a below normal summer, was 1969. Precipitation at Resolute Bay in July and August of 1969 was 160% of normal.

It is now necessary to determine whether conditions such as a cold wet summer combined with a normal or warm winter do occur, and to determine whether they are capable of producing observed values of δ and PC. We are limited to mass balance and δ measurements for the period 1962 to 1972. Fortunately, within this period several interesting extreme seasons occurred (Table 1).

1964 resembles the conditions hypothesized for the 1550-1620 period: warm δs, combined with low PC (4%) in a season with extreme positive mass balance conditions on all Queen Elizabeth Island (QEI) ice caps, July temperatures 2.2°C below normal in the QEI, and hemispheric cooling - particularly at high latitudes. For comparison, the characteristics of three other seasons should be noted.

In 1972 the coldest δ values combine with low melt values, hemispheric cooling and positive mass balance conditions on the QEI ice caps. In 1969 on Devon Island Ice Cap, the warmest δ values combine with high melt and negative mass balance conditions, while hemispheric conditions are near normal in high latitudes. In 1962, cool δs combine with maximum melt, near normal hemispheric conditions and highly negative mass balance on all Queen Elizabeth Island ice caps.

EVALUATING THE EFFECT OF THE RATIO BETWEEN SUMMER AND WINTER ACCUMULATION ON MEAN ANNUAL δ VALUES

Precipitation is not distributed evenly throughout the year. The actual temperature of snow accumulating on the ice cap, here referred to as the *accumulation weighted temperature* T_b, can be expressed as:

$$[1] \qquad T_b = \sum_{i=1}^{n} T_i \times PP_i$$

where T_i is the air temperature during each precipitation event (assumed to be linearly related to condensation temperature), and PP_i is the percent of the total annual precipitation that fell during that event. The accumulation weighted temperature (T_b) is

TABLE 1: *SEASONAL VALUES OF VARIOUS PARAMETERS FOR DEVON ISLAND ICE CAP, NORTHWEST TERRITORIES.*

YEAR	PC[1] (%)	$\delta 0^{18}$[2] (0/00)	MASS BALANCE[3] (kg m^{-2} year^{-1})	PERCENTAGE OF TOTAL OPEN WATER[4]	HEMISPHERIC TEMPERATURE DEVIATION[5] LATITUDE	
					(65-95^{0})	(25-90^{0})
1961	27	-27.7	-197	59	+	+
1962	56	-27.3	-359	70	0	+
1963	2	-27.1	+ 44	53	-	+
1964	4	-26.9	+125	34	--	-
1965	1	-27.2	+ 64	35	0	-
1966	31	-29.5	-135	46	--	-
1967	9	-31.0	+ 25	-	-	-
1968	0	-27.2	+ 5	35	--	--
1969	46	-26.5	-332	41	0	--
1970	8	-27.8	+ 39	37	-	--
1971	21	-29.1	- 69	50	--	--
1972	0	-29.1	+102	21	--	--

[1] PC - percentage of ice layers in snow between 1600 and 1800 m asl (after R.M Koerner 1977).

[2] $\delta 0^{18}$ - after D.A. Fisher (personal communication) error = 0.8^{O}/oo.

[3] Mass Balance - after Koerner (1977).

[4] Open Water - after Koerner (1977).

[5] Hemispheric Temperature Deviation - from 500/1000 mb thickness calculations after Dronia (1974) for 1949-1973 (-- very negative).

higher than the mean annual temperature at High Arctic locations because more precipitation falls during the warm season than during the cold season. This would not complicate the δ-T relationship if the distribution of precipitation and temperature remained the same from year to year and period to period.

Koerner (personal communication) has measured, along with the mass balance survey (usually in May), the percentage of the total annual accumulation which fell between the end of the previous melt season and the time of the May survey. For our purposes, this is referred to as the fall and winter accumulation (PP_w). These values are listed in Table 2. It can be seen that in 1964 (warm δ, low PC), the percent of fall and winter accumulation is a minimum, while in 1972 (cold δ, low PC) and 1979, fall and winter accumulation is a maximum. There is a 40% difference in the amount of accumulation deposited in spring and summer between these extremes. Evaluation of T_b for these extreme years and other years for comparison is the next step.

No year-round meterological records are available from the Devon Island Drill Site or elsewhere on Devon Ice Cap. Koerner (personal communication) observed that 50% of the fall and winter snow fell by the end of September. Equation [1] has been simplified to:

$$[2] \qquad T_b = \sum_{j=1}^{4} T_j \times PP_j$$

where j represents four accumulation periods based on Koerner's observations and the normal precipitation distribution at Eureka (Table 3). The values PP_j for any year can then be evaluated from Koerner's fall and winter accumulation percent values (PP_w). Examples and description of the symbols are given in Table 3.

The mean (T_j) temperature for each period was obtained as follows:

(1) The temperature records of stations in the vicinity of Devon Island Ice Cap having a year-round recording program were examined. The three-year (1954-56) temperature record from Site 2 on the Greenland Ice Cap was chosen as best representing the Devon Island Ice Cap Drill Site - particularly in terms of temperature range. The Site 2 station is at a similar latitude and altitude as the Devon Island Drill Site, although it is on the opposite side of Baffin Bay. Mean annual temperature at Site 2 is -24°C, whereas the accepted value for Devon Island Drill Site is -23°C.

(2) Site 2 temperatures were then compared to Resolute and Eureka mean monthly temperatures for each of the three years. The shape of the Eureka mean monthly temperature curve resembled Site 2 better than did the Resolute curve.

TABLE 2: *PERCENT FALL AND WINTER ACCUMULATION (PP_w) (R.M. KOERNER, PERSONAL COMMUNICATION), SIMULATED ICE CAP MEAN ANNUAL TEMPERATURE (T_a), CALCULATED ACCUMULATION TEMPERATURE (T_b), DIFFERENCE BETWEEN T_b AND T_a, AND OBSERVED SEASONALS (δ_o) (FISHER, PERSONAL COMMUNICATION). TEMPERATURES ARE CALCULATED FROM 11 AUGUST OF PRECEDING YEAR TO 10 AUGUST OF THE YEAR NAMED, δ SEASONALS WERE MEASURED FROM WINTER MINIMUM TO WINTER MINIMUM. THE ERROR IN THE δ SEASONALS DUE TO LAYERS OF DRIFTED SNOW IS ± 0.8°/oo (D.A. FISHER, PERSONAL COMMUNICATION).*

Year	PP_w(%)[1]	T_a(°C)	T_b(°C)	T_b-T_a	δ_o(°/oo)
1964	50	-24.1	-17.1	7.0	-26.9
1970	61	-24.7	-18.5	6.2	-27.8
1969	63	-23.9	-18.4	5.5	-26.5
1977	69	-24.3	-19.3	4.9	-
1961	65	-24.3 (1955)[2]	-19.5	4.8	-27.1
1962	85	-24.0 (1956)[2]	-20.9	3.1	-27.3
1962	85	-23.7 (1954)[2]	-21.3	2.4	
1972	86	-25.8	-21.2	4.0	29.1
1979	90	-25.6	-23.8	1.8	-

[1] The table is ordered from lowest to highest percent fall and winter accumulation (PP_w).

[2] Using Site 2 temperatures for a synoptically similar year (in brackets).

TABLE 3: *BREAKDOWN OF SEASONAL ACCUMULATION DISTRIBUTION USED IN THIS STUDY.*

MASS BALANCE SEASON		END OF MELT	FALL AND WINTER PP_w[1]		MID-MAY SURVEY	SPRING AND SUMMER PP_s[2] $= 100 - PP_w$	END OF MELT
ACCUMULATION PERIODS	j[3]	2/3 Aug. Sept.	Oct.	Nov. Dec. Jan. Feb. Mar. Apr. 1/2 May		1/2 May June July 1/3 Aug.	
EUREKA	PP_j[4]	.28	.10	.27		.35	
TO CALCULATE	PP_j[5]	$.47xPP_w$	$.15xPP_w$	$.38xPP_w$		PP_s	
DEVON MEAN	PP_j[6]	.31	.10	.25		.33	

[1] PP_w - fall and winter precipitation percent; % of total accumulation since end of past melt season measured at the time of the mass balance survey.

[2] PP_s - spring and summer precipitation.

[3] j - breakdown of year into four accumulation periods.

[4] Eureka PP_j - percent of total annual precipitation in each period from Eureka monthly precipitation normals.

[5] To calculate PP_j - precipitation percent formula for accumulation periods based on Eureka normal distribution and Koerner's observations(personal communication).

[6] Devon mean PP_j - precipitation percent for accumulation periods calculated using average observed $PP_w = .67$ for Devon Island Ice Cap (R.M. Koerner, personal communication).

(3) A correction was then calculated based on the mean values for the three years, to adjust the Eureka temperature to simulate conditions at Site 2.

(4) Thus, using the Eureka mean monthly temperatures for any year and the correction for each month, a simulated mean monthly temperature can be obtained for the Devon Island Drill Site in that year.

(5) From this the mean temperature for each accumulation period (T_j) in any specific year can be simulated.

(6) In addition, a simulated Devon Island Drill Site mean annual temperature (T_a) for particular years can be obtained (Table 2).

The "year" in this study represents the mass-balance year on Devon Island Ice Cap and runs from the end of the first 10 days of August (end of the melt season on Devon Island Ice Cap) to the same time the following season. The year is named by the melt season (i.e., 11 August 1971 to 10 August 1972 is called 1972).

The results of calculating accumulation-weighted temperature (T_b) from equation [2] for various seasons are given in Table 2. It can be seen that T_b is strongly dependent on the season of snow accumulation. There is a 6.7°C difference in T_b from 1964 to 1979, but only a 1.5°C difference in mean annual temperatures (simulated temperatures for Devon Island Drill Site). This means that 5.2°C of the difference was due to the 40% difference in ratio of spring and summer to fall and winter accumulation between these two years. This accumulation distribution difference in the mean annual temperatures ΔT_{b-a} can be expressed as follows (a list of symbols accompanies Table 4):

$$[3] \quad \Delta T_{b-a} = (T_{b^1} - T_{b^2}) - (T_{a^1} - T_{a^2})$$

where 1 and 2 refer to the year, and T_a is the simulated Devon Island Drill Site mean annual temperature. If Dansgaard's relationship $\delta = .62T$ (Dansgaard *et al.* 1973) is accepted, then ΔT_{b-a} can be converted to $\Delta\delta_{b-a}$, which represents the calculated accumulation distribution difference in δ seasonals between season 1 and season 2.

An equation similar to [3] can be written to obtain $\Delta\delta_{o-a}$, the accumulation distribution difference in observed δ seasonals (δ_o) between two years 1 and 2:

TABLE 4: *COMPARISON OF ΔT_{b-a} AND $\Delta\delta$ BETWEEN THE 1964 SEASON AND THE 1979 AND 1972 SEASONS.*

	ΔT_{b-a} (° C)	$\Delta\delta_{b-a}$ (°/oo)	$\Delta\delta_{o-a}$ (°/oo)
1964 - 1979	5.2	3.2	—
1964 - 1972	2.4	1.5	1.1

Key:

Standard deviation (σ) of δ seasonals is 0.8 °/oo (D.A. Fisher, personal communication).

T_j	is the mean temperature of an accumulation period (see Table 3).
T_b	is the accumulation weighted temperature.
T_a	is the simulated annual temperature.
δo	is the observed δ seasonals (D.A. Fisher, personal communication).
ΔT_b	is the difference in accumulated weighted temperature between two years.
$\Delta\delta_b$	is thus ΔT_b expressed in °/oo using Dansgaard's relationship $\delta = .62T$ (Dansgaard *et al.* 1973).
ΔT_{b-a}	is the temperature difference between two years, due only to differences in the accumulation distribution (i.e. with the temperature effect removed).
$\Delta\delta_{b-a}$	is ΔT_{b-a} expressed in °/oo.
$\Delta\delta_{o-a}$	is the observed δ difference, between two years, due only to differences in the accumulation (i.e. with the temperature effect removed).

$$[4] \quad \Delta\delta_{o-a} = (\delta_{o^1} - \delta_{o^2}) - .62\ (T_{a^1} - T_{a^2})$$

Comparing $\Delta\delta_{o-a}$ with $\Delta\delta_{o-a}$ (Table 4) for the extreme seasons 1964 and 1972, shows that calculated and observed values agree within the accuracy of the observations. No observed δ seasonals are available for 1979, but calculations show an accumulation distribution difference ($\Delta\delta_{o-a}$) of over 3 ^{o}/oo between 1964 and 1979.

QUANTITIVE COMPARISON OF 1964 AND THE PERIOD 1550-1620

As represented in Figure 2, the δ-PC relationship of 1964 corresponds to that of the 1550-1620 period in the core, while the 1972 relationship resembles the two more recent PC minima. The average δ values for the 1550-1620 period are about 0.6^o/oo less negative (warmer) than those accompanying the two most recent PC minima. The observed seasonal δ for 1964 is 2.2 ^{o}/oo less negative (warmer) than that for 1972. Caution must be exercised in comparing individual seasons with longer periods. However, using synoptic analogues for the QEI, it was seen that summer temperature differences between 50-year extreme periods were as much as .36 of those between extreme seasons (Alt 1983). Comparing the observed differences in core values between the two periods (0.6^o/oo) to the observed difference between δ seasonals for 1964 and 1972 (2.2^o/oo) gives a ratio of .27, which is well within these limits. This suggests the extreme years of 1964 and 1972 do quantitatively represent the differences seen in the core values.

To determine whether accumulation distribution alone accounts for the 0.6 ^{o}/oo difference between 1550-1620 and the more recent PC minima, reference should be made to the values in Table 4. Between 1964 and 1972, roughly half of the difference in the δ values was due to the differing accumulation distribution. Between 1964 and 1979, however, the accumulation distribution difference was 80% of the total difference. In the latter case accumulation distribution difference in the δ signal was more than five times greater than the 0.6^o/oo difference seen in the core. If, as seems most likely, the two most recent PC minima were dominated by seasons such as 1972, the δ difference between these periods and the 1550-1620 period was equally a result of warmer mean annual temperatures and accumulation distribution differences. Periods dominated by situations such as occurred in 1979 would be expected to experience strongly negative (cold) δs due more to precipitation distribution than temperature.

As seen in Alt (1983), mean annual temperatures at QEI stations were higher in 1964 than in 1972, while summer temperatures in 1964 were lower than in 1972.

To summarize, quantitative considerations suggest that the 1550-1620 period had a 30 to 40% greater proportion of accumulation (snow) in the spring and summer than the two most recent PC minima, and that this combined with warmer winter temperatures to produce less negative δ values in the core.

THE PERIOD 1550-1620 AS SIMULATED BY 1964

Periods in the ice-core records that exhibit the characteristics of the reference seasons (1964, 1972 and 1969) have been indicated on the time series plots of $\delta 0^{18}$ and melt percent (Figure 3). Schematic representations of the synoptic conditions that dominate the summers of 1964, 1972 and 1969 are also included.

The 1964 synoptics show frequent cyclonic activity. Low pressure systems extending from the surface into the upper troposphere track into the Queen Elizabeth Islands from the north and northwest, having travelled for at least four days over the Polar Ocean (Alt, in preparation). These systems contain below-freezing temperatures and snow. Melt does occur between systems, or when warm air from the south intrudes north, but snow deposited by the frequent Polar Ocean lows maintains high surface albedos, thus limiting the melt. Highly positive mass-balance conditions are then experienced on all Queen Elizabeth Island ice caps. In the case of 1964, this was the result of enhanced winter precipitation as well as summer accumulation (Alt, in preparation).

The period 1550-1620 is preceded by high δ and high PC values such as those in 1969 on Devon Island Ice Cap. The 1969 synoptics show frequent cyclonic activity but, in contrast to the 1550-1620 period (represented by 1964), the surface systems track from southwest around an upper atmosphere system located in the Beaufort Sea sector of the Polar Ocean. The Queen Elizabeth Islands are frequently under the influence of the warm sector resulting in high melt. In the northern region of the QEI snow falls between warm sector intrusions.

Two periods of coinciding δ and PC minima (approximately 1650-1720 and 1790-1860) are thought to resemble 1972. Hemispheric cooling, winter conditions extending into the melt season, little melt and average accumulation maintains a positive mass balance on the ice caps. The more extreme case of 1979 will also be examined in connection with a detailed study of the 1800-1860 period.

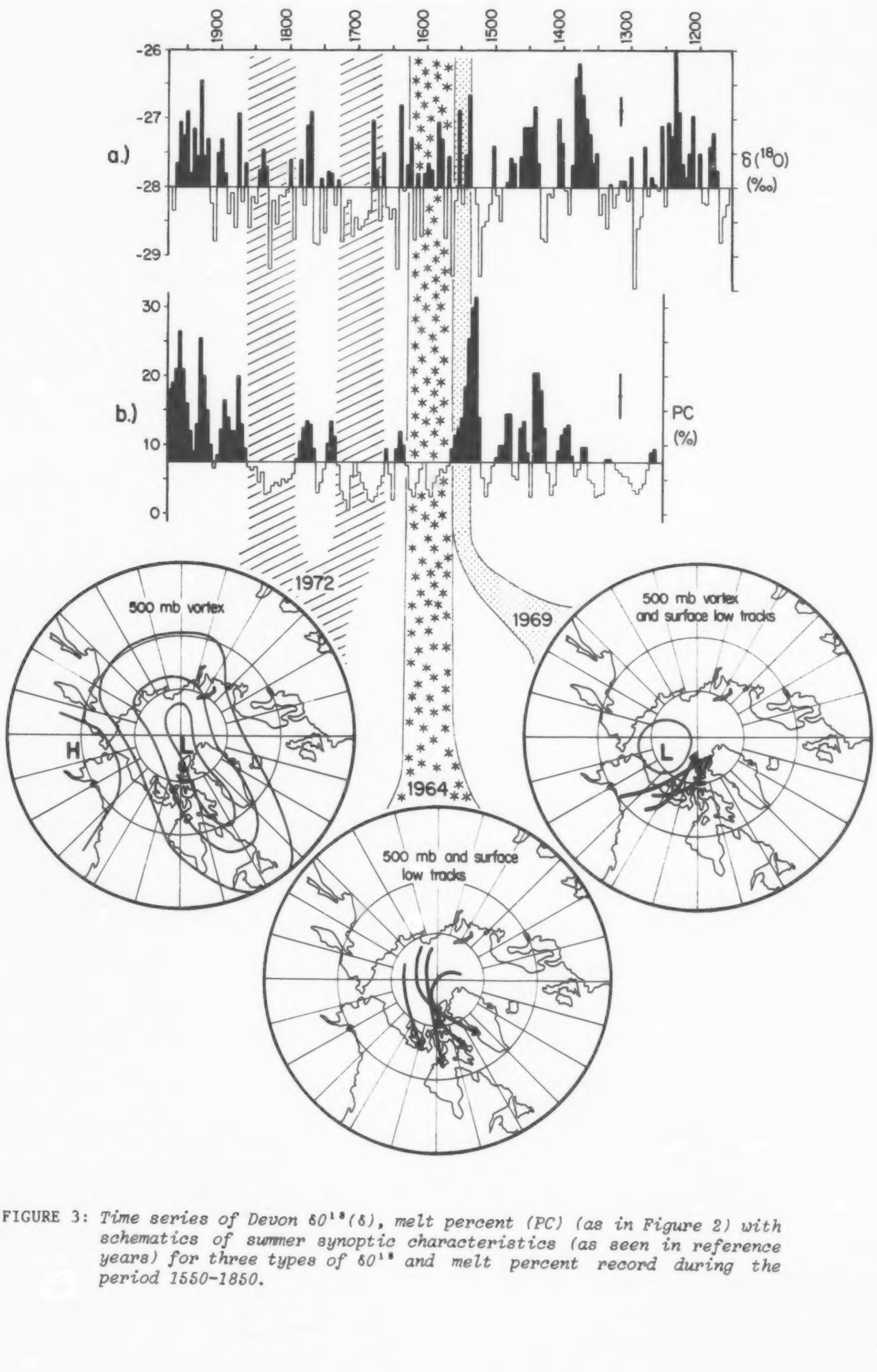


FIGURE 3: *Time series of Devon $\delta O^{18}(\delta)$, melt percent (PC) (as in Figure 2) with schematics of summer synoptic characteristics (as seen in reference years) for three types of δO^{18} and melt percent record during the period 1550-1850.*

A period of low δ and high PC (about 1860-1890) when melt begins to rise while remains low, may resemble 1962, where a ridge pushes into the Queen Elizabeth Islands often enough to produce strongly negative mass balance conditions without raising hemispheric temperatures significantly.

GENERAL ATMOSPHERIC CIRCULATION CHARACTERISTICS: 1550-1620

Queen Elizabeth Island synoptic studies (Alt 1983) suggest that climatic change events in the QEI result from hemispheric cooling or a shift in the position of the circumpolar vortex. Based on similarities to the reference years, the Little Ice Age in the Queen Elizabeth Islands can be seen as a period of hemispheric cooling augmented by two distinct positions of the circumpolar vortex.

During the two most recent melt minima (reference year 1972) elongation and deepening of the North American trough resulted in a long deep trough from Labrador-Ungava across the Pole to the Kara-Laptev Sea (Figure 3).

Prior to 1550, rain and high melt were experienced in the southern Queen Elizabeth Islands in summer as cyclonic systems with well-defined warm sectors entered the QEI from the southwest (reference year 1969). By about 1550 a ridge builds into the Mackenzie area of North America (perhaps linked with a shift of the stratospheric circumpolar vortex into the European sector in early spring (Alt, in preparation)), and surface and upper tropospheric lows track into the QEI from the Polar Ocean (reference year 1964). Moisture from the Laptev, Kara and perhaps even Norwegian and Barents seas is augmented by that picked up from leads and melting snow on the Polar Ocean. On reaching the mountains of the eastern Queen Elizabeth Islands, the moisture falls as summer snow on QEI ice caps. The air mass contains near-freezing temperatures, even at sea level during July. Winter accumulation may also have been increased. The Queen Elizabeth Island ice caps would be expected to expand. As δs at Camp Century on western Greenland (Fisher 1982) are not above the 800-year average, this area is thought to experience cooling but not the summer accumulation. In eastern Greenland (Crete) and Iceland the cooling is absent, while still farther east, in England, cooling also begins about 1550 (Dansgaard *et al.* 1975).

From these considerations and the 1964 synoptic conditions, the summer long-wave pattern during the 1550-1620 period may be depicted as shown in Figure 4: (a) weak ridges in the Mackenzie-Keewatin area, over eastern Greenland and the Barents Sea, and (b) strong troughs

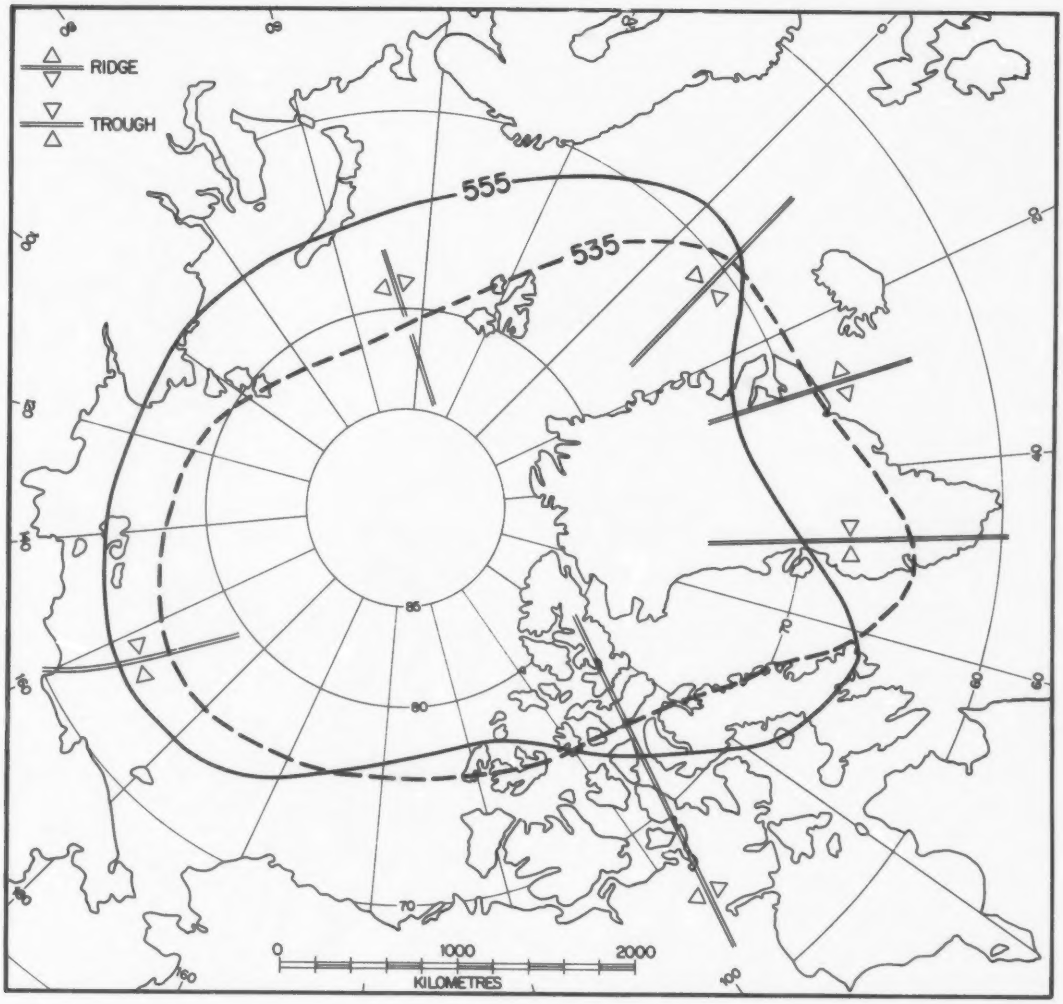

RIDGE
TROUGH
555
535
85
80
70
0
1000
2000
KILOMETRES

FIGURE 4: *Schematic comparison of long-wave pattern (500 mb heights in dekameters) for present (solid line) and the 1550-1620 period (dashed line).*

down western Greenland, between Greenland and Spitzbergen to the British Isles and in the East Siberian Sea. A shift eastward of the North America trough is particularly evident in the schematic representation of Figure 4.

CONCLUSIONS

Details of this synoptic climate recreation for the initial period of the Little Ice Age in the Queen Elizabeth Islands (1550-1620), may be augmented or altered by examination of other ice-core parameters (e.g. electrolytic conductivity, pollen). The method shows that analysis of one proxy data parameter on its own may be misleading. On the other hand, examination of several parameters for a specific period provides an understanding of climatic change, not as the trend of a single parameter, but as a complex interaction of many variables on various time and space scales at various levels in the atmosphere and oceans. As we assemble proxy data from various latitudes and longitudes, we must begin to piece together the response of regional climates to the long-wave patterns in the atmosphere.

ACKNOWLEDGEMENTS

R.M. Koerner and D.A. Fisher made available unpublished data and provided much needed advice on the use of these and other ice-core data. They also made extensive comments on the first draft of the paper.

REFERENCES

Alt, B.T. 1980. Synoptic climate factors governing extreme mass balance seasons on Queen Elizabeth Island glaciers: (1960-1978). Implications for paleoclimatic studies. Final Report Contract OSQ77-00239 for Polar Continental Shelf Project, Ottawa. 500 pp.

______. 1983. Synoptic analogs: a technique for studying climatic change in the Canadian High Arctic. In: Climate Change in Canada 3. Edited by: C.R. Harington. Syllogeus No. 33:70-107.

______. (In preparation). Developing synoptic analogs for extreme mass balance conditions in the Queen Elizabeth Islands.

Dansgaard, W., S.J. Johnsen, H.B. Clausen, and N. Gundestrup. 1973. Stable isotope glaciology. Meddelelser om Grønland 197:1-53.

Dansgaard, W., S.J. Johnsen, N. Reeh, H.B. Gundestrup, and C.U. Hammer. 1975. Climatic changes, Norsemen and modern man. Nature 255 (5503):24-28.

Dronia, H. 1974. Über Temperaturänderungen der freien Atmosphäre auf der Nordhalbkugel. In: In den letzten 25 Jahren. Meteorologische Rundschau 27:166-174.

Fisher, D.A. 1977. Melt feature stratigraphy, oxygen isotope ratio, and climate over 800 years from the Devon Island ice cap cores, a final report. Polar Continental Shelf Project (internal report). 33 pp.

______. 1982. Carbon-14 production compared to oxygen isotope records from Camp Century, Greenland and Devon Island, Canada. Climatic Change 4:419-426.

Fisher, D.A., and R.M. Koerner. 1981. Some aspects of climatic change in the High Arctic during the Holocene as deduced from ice cores. In: Quaternary Paleoclimate. Edited by: W.C. Mahaney. Geo Books, University of East Anglia, Norwich. pp. 249-271.

______. 1983. Ice-core study: a climatic link between the past, present and future. In: Climatic Change in Canada 3. Edited by: C.R. Harington. Syllogeus No. 49:50-69.

Koerner, R.M. 1977. Devon Island Ice Cap: core stratigraphy and paleoclimate. Science 196 (4285):15-18.

Koerner, R.M., and D.A. Fisher. 1981. Studying climatic change from Canadian High Arctic ice cores. In: Climatic Change in Canada 2. Edited by: C.R. Harington. Syllogeus No. 33:195-215.

Paterson, W.S.B., R.M. Koerner, D. Fisher, S.I. Johnsen, H.B. Clausen, W. Dansgaard, P. Bucher, and M. Oeschger. 1977. An oxygen-isotope climatic record from the Devon Island Ice Cap, Arctic Canada. Nature 266 (5602):508-511.

HOW DID WE GET INTO THE LAST GLACIATION, HOW DID WE GET OUT OF IT, AND WHAT HAPPENED IN BETWEEN?

C.U. Hammer[1]

ABSTRACT

Data on oxygen isotopes, acidities and dust-concentrations from Greenland, Antarctic and Canadian deep-ice cores present a pattern of climatic events, which can be summarized as follows:

(1) The Holocene shows higher oxygen-isotope ratios, than the Wisconsin glaciation; acidities are in the range of 0.6-3 μ equiv. H^+/kg (depending on location and time) and dust-concentrations are generally 5-20 times lower in Holocene as compared to Wisconsin ice.

(2) The Greenland Ice Sheet is generally alkaline over its Wisconsin ice, while the corresponding Antarctic ice is acid.

(3) Climate of the Wisconsin glaciation differed in the southern and northern hemisphere. A number of rapid and drastic climatic changes took place in the northern hemisphere during the last glaciation, while such changes were less rapid and less drastic in the southern hemisphere. The Allerød and Younger Dryas climatic epochs seem to be northern hemisphere phenomena.

Analysis of new data from the Dye 3 deep core in southern Greenland offers a descriptive explanation of how glacial climate changed to Holocene conditions. Climatic variations during the ice age must be closely linked to glacial retreats and advances.

The "jump" into the last glaciation, or the "jump" into a future glaciation may only need a small triggering effect. The possibility of volcanism as such a triggering effect is worth discussing.

---

[1] Geofysisk Isotop Laboratorium, Københavns Universitet, Haraldsgade 6,2200 København N.

RECENT SYLLOGEUS TITLES / TITRES RECENTS DANS LA COLLECTION SYLLOGEUS

No. 42 Shih, Chang-tai, and/et Diana R. Laubitz (1983)
SURVEY OF INVERTEBRATE ZOOLOGISTS IN CANADA - 1982 / RÉPERTOIRE DES ZOOLOGISTES DES INVERTÉBRÉS AU CANADA - 1982. 93 p.

No. 43 Ouellet, Henri et Michel Gosselin (1983)
LES NOMS FRANÇAIS DES OISEAUX D'AMÉRIQUE DU NORD. 36 p.

No. 44 Faber, Daniel J., editor (1983)
PROCEEDINGS OF 1981 WORKSHOP ON CARE AND MAINTENANCE OF NATURAL HISTORY COLLECTIONS. 196 p.

No. 45 Lanteigne, J. and D.E. McAllister (1983)
THE PYGMY SMELT, *OSMERUS SPECTRUM* COPE, 1870, A FORGOTTEN SIBLING SPECIES OF EASTERN NORTH AMERICAN FISH. 32 p.

No. 46 Frank, Peter G. (1983)
A CHECKLIST AND BIBLIOGRAPHY OF THE SIPUNCULA FROM CANADIAN AND ADJACENT WATERS. 47 p.

No. 47 Ireland, Robert R. and Linda M. Ley (in press)
TYPE SPECIMENS OF BRYOPHYTES IN THE NATIONAL MUSEUM OF NATURAL SCIENCES, NATIONAL MUSEUMS OF CANADA.

No. 48 Bouchard, André, Denis Barabé, Madeleine Dumais, et/and Stuart Hay (1983)
LES PLANTES VASCULAIRES RARES DU QUÉBEC. / THE RARE VASCULAR PLANTS OF QUÉBEC. 79 p. / 75 p.

No. 49 Harington, C.R., editor (1983)
CLIMATIC CHANGE IN CANADA 3. 343 p.

No. 50 Hinds, Harold R. (1983)
THE RARE VASCULAR PLANTS OF NEW BRUNSWICK. / LES PLANTES VASCULAIRES RARES DU NOUVEAU-BRUNSWICK. 38 p. / 41 p.

No. 51 Harington, C.R. and G. Rice, editors; Anne B. Smithers, Linda Ghanimé and C.R. Harington, compilers (1984)
CLIMATIC CHANGE IN CANADA 4, ANNOTATED BIBLIOGRAPHY OF QUATERNARY CLIMATIC CHANGE IN CANADA. 368 p.

No. 52 Hunter, J.G., S.T. Leach, D.E. McAllister and M.B. Steigerwald (1984)
A DISTRIBUTIONAL ATLAS OF RECORDS OF THE MARINE FISHES OF ARCTIC CANADA IN THE NATIONAL MUSEUMS OF CANADA AND ARCTIC BIOLOGICAL STATION. 35 p.

No. 53 Russell, D.A. (1984)
A CHECK LIST OF THE FAMILIES AND GENERA OF NORTH AMERICAN DINOSAURS. 35 p.

No. 54 McAllister, Don E., Brad J. Parker and Paul M. McKee (1985)
RARE, ENDANGERED AND EXTINCT FISHES IN CANADA. 192 p.